

Thesis Submitted for the Degree of
Doctor of Philosophy at the
University of Oxford

PETROGENETIC STUDY OF THE GLAUCOPHANE SCHISTS
AND ASSOCIATED ROCKS FROM
THE ILE DE GROIX, BRITTANY, FRANCE

by

Michael S.N. Carpenter (B.A. Oxon)

of

Brasenose College, Oxford
and the Department of Geology and Mineralogy,
Oxford.

November 1976



TABLE OF CONTENTS

	<u>Page</u>
Acknowledgements	i
Abstract	iii
CHAPTER ONE - INTRODUCTION	
1.1	1
1.2 Aims of the present study	3
1.3 Outline of regional geology	4
1.4 General description of the Ile de Groix - fieldwork technique	7
CHAPTER TWO - CLASSIFICATION AND DESCRIPTION OF ROCK TYPES EXPOSED ON THE ILE DE GROIX	
Introduction	10
(1) The Aluminous Schists	13
IA Aluminous glaucophane schists	13
IB Aluminous chloritoid schists	15
IAB Aluminous glaucophane - chloritoid	17
IC Aluminous non-glaucophane, non-chloritoid schists	18
a) Albitic micaschists	18
b) Quartzose micaschists	21
(2) The Banded Epidote Schists	23
Chemical and paragenetic variation of the banded epidote schists	24
a) Epidote + glaucophane + garnet $\pm$ phengite $\pm$ Qz $\pm$ Chl $\pm$ Rutile $\gg$ Sphene	30
b) Epidote + glaucophane + phengite $\pm$ Chl $\pm$ Qz	31
c) Epidote + glaucophane + chlorite + albite $\pm$ calcite	33
d) Epidote + glaucophane + chloritoid + garnet $\pm$ alb $\pm$ paragonite $\pm$ phengite	35

	<u>Page</u>
e) Epidote + garnet + chloritoid + albite ± phengite	37
f) Epidote + chlorite + albite ± ilm + sphene	38
(3) <u>Magnesian Schists</u>	41
(4) Massive Basic Schists	43
a) Pyroxene bearing glaucophane garnet schists or glaucophanic eclogites	43
b) Chlorite + albite greenstones	49
c) Actinolite + albite + chlorite + epidote	53
CHAPTER THREE - STRUCTURAL AND TEXTURAL SYNTHESIS	
3.1 Introduction	58
3.2.1 S ₀ Original Compositional Layering	59
3.2.2 F _e 'deformation'	60
3.2.3 S ₁ schistosity	60
3.2.4 F ₁ deformation	60
3.2.5 F ₂ deformation	61
3.2.6 F ₃ deformation	67
3.2.7 F ₄ deformation	69
3.3 Geographical distribution of tectonic style	81
3.4 Cogné's interpretation of the structuro-metamorphic history of Groix	74
3.5 Objections to Cogné's model	76
3.6 Mineral growth episodes in relation to deformational history	80
3.6.1 Evidence from rutile/sphene relationships for a prograde (post F ₁) development of garnet + omphacite assemblage in dehydrated zone I metabasics	82
3.6.2 Evidence for primary syntectonic growth of glaucophane	83
3.6.3 The relationship of glaucophane to garnet and omphacite	84

	<u>Page</u>
3.6.4 Evidence for post-tectonic garnet growth	88
3.6.5 Evidence for glaucophane + lawsonite paragenesis during F_1 or F_2	89
3.6.6 Augen structures do not necessarily imply pre F_1 growth of garnet omphacite assemblage	92
3.6.7 Early garnet growth is syn lawsonite breakdown but pre sphene growth in Zone I metabasics	92
3.7 Evidence for partial and complete retrogression post F_2 deformation phase	94
3.7.1 Partial retrogression	95
3.7.2 Complete retrogression of metabasic schists	103
CHAPTER FOUR - CONDITIONS OF METAMORPHISM	
4.1 Introduction	106
4.2 Schreinemaker's analysis of stable phase assemblages	110
4.3 Example 1	111
4.4 7 phase, 4 component multisystem for metabasic parageneses on Groix	116
4.5 Comparison of model petrogenetic grid relations with natural metabasic assemblages on Groix	122
4.6 Conclusions drawn from comparison of model petrogenetic grids and natural metabasic assemblages	124
4.7 Influence of grade and host rock chemistry on mineral chemistry	127
4.8 Estimation of pressure, temperature and fluid phase composition during metamorphism of Groix metabasics	131
4.9 Application of Raheim and Green's (1974) Fe/Mg partition geothermometer	132
4.10 Extrapolation of Raheim and Green's geothermometer to temperatures less than 600°C.	141

	<u>Page</u>
4.11 Estimation of load pressure ($= P_{TOTAL}$) during metamorphism of the Groix glaucophanic eclogites	145
4.12 Jadeite + Quartz $\lesseqgtr$ Albite equilibrium	147
4.13 Recalculation of the pyroxene formulae	151
4.14 Activity-composition relationship for jadeite in multicomponent solid solution	153
4.15 Comparison of Ganguly's model with natural paragenetic data	156
4.16 Estimation of fluid phase composition during metamorphism	159
4.17 CO ₂ content of the fluid phase attending metamorphism of the zone I and zone II Groix metabasics	162
4.18 Further checks on the pressure, temperature and fluid phase composition during prograde metamorphism on Groix	165
CHAPTER FIVE - GEOCHRONOLOGY OF THE ILE DE GROIX	
5.1 Introduction	171
5.2 Strontium isotope and trace element studies on the metabasic schists, and some associated rocks, from the Ile de Groix	173
5.2.1 Trace element characterization of magma types	177
5.2.2 Comparison of trace element geochemistry of metabasics with Sr isotope data	178
5.2.3 Estimation of the maximum age of eruption of the Groix metabasics	182
5.3 K-Ar dating of metamorphic events in the Ile de Groix blueschists	186
5.4 Rb-Sr mineral isochron for the glaucophanic eclogite, and model ages for phengites from the Ile de Groix	191
5.5 Summary statement	195

	<u>Page</u>
CHAPTER SIX - CONCLUSION	
6.1 Introduction	198
6.2 Synthetic geological history postulated for the Ile de Groix blueschists	198
6.3 Problem of water supply during retrogression	200
6.4 Suggestions for the further testing of hypotheses given in this study	201
Appendix I Sample descriptions	204
Appendix II Whole rock chemistry	210
Appendix III Electron probe analyses of minerals from the Ile de Groix	224
Appendix IV Graphical presentation of mineral analyses	256
References	263

LIST OF FIGURES

CHAPTER ONE

1.1 Position of the Ile de Groix relative to major Hercynian shear belts	2
1.2 Cogné's (1960) postulated relationship of Variscan granite 'lobes' to a shear belt on the S.W. margin of the Anticlinal de Cornouailles.	5

CHAPTER TWO

2.1 Sample localities and distribution of prograde metamorphic zones	9
2.2 NA-C-A-FM projection from albite of Groix epidote banded schist parageneses	26

2.3	Compositional and grade control on banded epidote and massive basic schist parageneses	28
2.4	Highly sheared, epidote banded schist with flat lying F_1 isoclinal	32
2.5	There is no Figure 2.5	
2.6	Field relationships at the Crique de Biléric locality	40
2.7	F_4 fold in chlorite + albite greenstone band	42
2.8	Contact between massive glaucophanic eclogite and banded epidote schist horizon	45
2.9	NA-C-A-FM projection from albite of Groix metabasic parageneses	48
2.10	Gradation between glaucophane-rich banded metabasic schists and cavernous weathering greenstones	52
2.11	PHOTOMICROGRAPH (x 45) of spec. BIL3	54
CHAPTER 3		
3.1	Schematic cross section of the south coast of the Ile de Groix	57
3.2	Type a) F_1 isocline in banded epidote schists	62
3.3	Metabasic boudins encased in chloritoid micaschists	63
3.4	Interference 'fuseaux' in graphitic micaschists	65
3.5	Epidote rich bands pick out a refolded F_1 isocline	66
3.6	F_3 strain slip cleavage in micaschists	68
3.7	Stereographic projection	70
3.8	Model for the production of boudins in triaxial F_1 stress field	79
3.9	PHOTOMICROGRAPH (x 45) 2.3 mm euhedral porphyroblast of garnet with inclusion of ferroglaucophane. TUD5	86
3.10	A profuse development of pseudomorphs after lawsonite	91

	<u>Page</u>
3.11 PHOTOMICROGRAPH (x 45)	
PC10 pyroxene augen	87
3.12 PHOTOMICROGRAPH (x 72)	
Garnet euhedra in pseudomorph material. Spec. PM	93
3.13 PHOTOMICROGRAPH (x 45)	
Alteration of ferroglaucophane in KB1	95
3.14 PHOTOMICROGRAPH (x 45)	
Chloritoid + glaucophane alteration in PM10	98
3.15 PHOTOMICROGRAPH (x 45)	
Alteration of garnet to chlorite in PC1	99
CHAPTER FOUR	
4.1 Model petrogenetic grid for basic assemblages on Groix	113
4.2 Model petrogenetic grid for 7 phase, 4 component multi-system	119
4.3 Path of prograde metamorphism in Groix metabasics in relation to proposed petrogenetic grid	125
4.4 Comparison of K_D 's observed at constant pressure, with K 's calculated from mole fraction data.	138
4.5 Extrapolation of Raheim & Green's geothermometer to temperatures less than 600°C.	143
4.6 Pressure/temperature conditions of equilibration for garnet/omphacite pairs (this study) from Groix	148
4.7 Activity compositions relationship in system diopside-jadeite	154
4.8 T-fo ₂ stability field for ferroglaucophane	161
4.9 Summary diagram for proposed pressure/temperature conditions during prograde and retrograde metamorphism on Groix	170
CHAPTER FIVE	
5.1 Rb/Sr versus $^{87}\text{Sr}/^{86}\text{Sr}$ plot for Groix metabasics	176

5.2	$^{87}\text{Sr}/^{86}\text{Sr}$ initial versus Sr ppm assuming model age of 400 Myr	180
5.3	Compilation of $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr data on rocks from the Ile de Groix	183
5.4	$^{87}\text{Sr}/^{86}\text{Sr}$ evolution diagram for Groix metabasics	185
5.5	Rb/Sr mineral isochron for phengite-epidote pair from glaucophane eclogite PC10	193
5.6	Summary of plutonic, volcanic, sedimentation and metamorphic events in Brittany	197
APPENDIX I		
I.	Epidote rich bands folded into flat-lying F_1 isoclinal. Locality TRE2	206
APPENDIX II		
II.1	An AFM plot of the basic schists from the Ile de Groix	219
II.2	Histogram of TiO_2 contents in metabasics from the Ile de Groix	220
II.3	Ti-Zr-Y discriminant diagram for Groix metabasics	221
II.4	REE patterns from zone II Groix metabasics	222
II.5	REE patterns from zone I Groix metabasics	223
APPENDIX IV		
IV.1	Zoning paths in Groix garnets	257
IV.2	Triangular composition diagram for clinopyroxenes	258
IV.3	Partition of Fe/Mg between co-existing garnet and clinopyroxene	259
IV.4	Composition of glaucophanes	260
IV.5	Projection of co-existing mineral compositions from epidote	261
IV.6	Composition of micas	262

ACKNOWLEDGEMENTS

The research for this thesis was carried out during the tenure of a Natural Environment Research Council award, under the supervision of Dr. J.D. Bell at the Department of Geology and Mineralogy, Oxford.

Throughout the course of research, Dr. Bell has provided constant encouragement and advice, particularly during the writing up, when his professional and psychological support proved very valuable.

The author is also much indebted to the technical staff of the Department of Geology, Oxford, for their help in laboratory work. Special thanks are due to Mr. C. Fagg, for XRF instruction, Mr. R. Goodwin for advice on rock crushing and mineral separation, Messrs. R. McAvoy and S. Baker for photography, Mr. M. Hepher for performing CO₂ analyses and Mr. R. Holland for making thin sections.

Miss Pam Shreeve deserves a special mention for her invaluable atomic absorption analyses of various rocks and minerals, which greatly assisted the author's geochronological studies.

Drs. J.V.P. Long, P. Statham and Mr. G.G. Price of the Department of Mineralogy and Petrology, Cambridge gave freely of their time to help in the author's microprobe work, as did Dr. W.A.P. Nicholson during the probe work at Oxford.

During field work on the Ile de Groix, the present author greatly benefited from discussions with Drs. Tarney and Wright of Birmingham University, Adolphe Nicholas of Nantes University, and enjoyed the hospitality of M. Serge Bihan. Mr. J.W.B. Carpenter provided road transport, field assistance and many of the excellent original photographs.

Dr. R.A. Roach of Keele University and Dr. A.J. Barber of Chelsea College, and their research students H. Roper, L. Richards and S. Hanmer provided much stimulating discussion of the regional geology of Brittany during a Geological Association field trip (Easter, 1975).

Dr. P.N. Taylor, working at the Oslo geological museum, Norway, has very generously supplied the Sr isotope analyses quoted in this study. Mineral separates and thin sections from Mäkinen's (1967) thesis work were supplied by the kind permission of Prof. R.A. Howie of King's College, University of London.

Drs. L. Civetta and J-J. Peucat (of Naples and Rennes Universities respectively) must be greatly thanked for their K/Ar and Rb/Sr mass spectrometric analyses on separated minerals.

Amongst the author's post-graduate colleagues at Oxford, D.J. Waters, S.R. Carter and M.J. Norry, R. Cleverley, P. Betton and P. Webb have been particularly helpful in discussion, laboratory work and thesis preparation.

The patient, and fast, typing work of Miss Sarah Fraser and Mrs. Pat Jackson is gratefully acknowledged. Mr. P. Deussen very kindly undertook most of the diagram draughting.

ABSTRACT

The metamorphic rocks of the Ile de Groix are classified into four groups:

1. Aluminous schists
2. Banded epidote schists
3. Magnesian schists
4. Massive metabasic schists

which are then subdivided on the basis of observed mineral paragenesis.

Rocks of presumed igneous parentage (including the banded epidote schists, massive metabasic schists and some albitic aluminous schists) were selected for detailed chemical and mineralogical investigation, and twenty new whole rock analyses (major and trace elements) are given.

Of the twenty analysed rocks, eight provided material for original microprobe analyses of the constituent minerals and sixteen were analysed for their Sr isotopic composition. Co-existing phengite, glaucophane, epidote and impure barroisite were separated from one of the analysed rocks (a glaucophanic eclogite) and used to obtain K-Ar and Rb-Sr (mineral isochron) radiometric ages. This dating work was supplemented by the inclusion of original K-Ar analyses on phengites and glaucophanes from aluminous and epidote banded schists, separated and provided by previous workers. The phengite separates, for which published analyses are available, were also used to yield Rb-Sr model ages.

On the basis of mineral parageneses observed in metabasic schists, the Ile de Groix is divided into two metamorphic zones, I and II, separated by a 'garnet-in-metabasics' isograd. In zone I are confined the phyllitic chloritoid schists and glaucophane bearing metabasic schists without garnet. Zone II contains garnet bearing porphyroblastic chloritoid schists and boudins of glaucophanic eclogite.

Within the two metamorphic zones so defined, the sequence of mineral growth in contrasting lithologies is inferred from thin section textural

studies. This sequence of mineral growth is then related to the development of fabric elements within the rock, as deduced from outcrop scale observations of fold style and schistosity.

Four phases of deformation are defined, using the criteria of re-folded folds and schistositities, and these are used as a relative time framework against which mineral growth episodes are compared.

F_1 deformation, characterised by flat lying, highly flattened isoclinal and boudins, is dominant in the higher grade zone II, and F_2 , characterised by open, upright folding, is dominant in zone I. In the boundary zone between the two zones, fold interference is very common.

Glaucophane growth in the metabasic schists is inferred to be mostly syn F_1 deformation, though the lack of straining suggests mimetic crystallization. The possibility of a glaucophane + lawsonite paragenesis during early syn-deformational phase of metamorphism in zone II is strongly supported by the occurrence of pseudomorphs after lawsonite which contain euhedral, post-tectonic garnets. Omphacite is thought to have crystallized in water-deficient assemblages after the breakdown of lawsonite and the beginning of garnet growth. Thus, the garnet + omphacite equilibrium assemblages are not relict from a pre-glaucophane schist metamorphism, but represent the peak of dry, prograde metamorphism in the metabasic rocks of Groix.

The conditions of metamorphism for the Groix blueschists are inferred from the comparison of natural, microprobed parageneses with published experimental mineral stability and element partitioning data. Pressure/temperature estimates for the metamorphism of the glaucophanic eclogites are derived from the Fe/Mg partitioning between adjacent garnet and clinopyroxene rims and from the jadeite content of omphacite in equilibrium with quartz and albite. This method gives an intersection of equilibria at $400 \pm 50^\circ\text{C}$ and 8 ± 1.5 kb pressure, which is consistent with blueschist facies rocks from New Caledonia and Japan, and suggests that metamorphism took place in a region of low geothermal gradient (ca. $15^\circ\text{C}/\text{km}$ if density = 2.9 g.cm^{-3}) - possibly

in a subduction zone.

Ferroglaucophane is commonly observed in the metabasic schists of zone I, and sometimes as inclusions in zone II garnets. From the rather restricted known temperature/ f_{O_2} stability of ferroglaucophane, it is possible to say that during metamorphism of the zone I metabasics, temperatures never rose more than about $390 \pm 10^\circ\text{C}$ at oxygen fugacities on the QFM buffer near 10^{-30} bars.

The T- XCO_2 stabilities of lawsonite and rutile/sphene place a constraint of $\text{XCO}_2 < 0.03$ during prograde metamorphism of the basic rocks.

A 'retrogressive' phase of metamorphism is proposed, associated with the deformation phases F_3 and F_4 , and is characterised by increased f_{O_2} (10^{-20} bars) decreased load pressure, and possibly higher temperatures ($450 - 500^\circ\text{C}$). These conditions are derived from the alteration of ferroglaucophane to magnetite, albite and quartz and the lack of kyanite development in aluminous rocks. Other effects of the 'retrogression' are the replacement of garnet, omphacite and glaucophane by barroisite, the alteration of garnets to chlorite and the infilling of cavities and veins with pegmatitic chlorite, albite and quartz. By extension, the massive chlorite + albite greenstones are thought to have developed during the 'retrogression' whenever water was available in sufficient quantities to allow recrystallization of the whole rock mass.

The Sr isotope compositions and present day Rb/Sr values of sixteen analysed rocks were used to place constraints on the original age of eruption of the Groix volcanics, which is thought to lie between 400 and 500 Myr ago. Fourteen of the rocks so used are metabasic, and three original magma type groups are recognised on the basis of trace element contents and discriminant analysis. One magma-type is common amongst the metabasics of zone II, and is characterised by Mid-Ocean ridge type chemistry and bimodal TiO_2 contents. Other magma types have more alkaline affinities, and a large range in $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio 400 Myr ago is postulated (0.704 - 0.711).

The concordance of K-Ar dates for two glaucophane-phengite pairs from widely separated aluminous and basic schist outcrops with the Rb-Sr mineral isochron (Epidote-whole rock-phengite) for a glaucophanic eclogite is interpreted as resulting from short-lived prograde mineral growth 340 ± 10 Myr ago throughout Groix.

Barroisite overgrowths on glaucophane are found to be significantly younger by about 40 Myr than the prograde metamorphic assemblage. The age of this late 'retrogression' (295 ± 8 Myr) corresponds with uplift and cooling ages throughout the Hercynides, and was probably brought about by thermal relaxation to a more normal geothermal gradient (i.e. $30^{\circ}\text{C}/\text{km}$), during rapid uplift, after subduction ceased.

CHAPTER 1

INTRODUCTION

1.1.

Situated off the south coast of Brittany in France, 10 km out to sea from the estuary of the Blavet near Lorient, the Ile de Groix appears to belong to the southern margin of a complex metamorphic belt that extends over South Brittany from the Pointe du Raz south-eastwards across the Loire and into the Vendée (see Fig. 1.1).

The Ile de Groix has recently attracted geological attention because it possesses the sole recognised occurrence of glaucophane schist within the entire Hercynides of western Europe. As such, it stands in complete contrast to the rest of the Hercynides, where metamorphic facies are generally low grade and related to the observed voluminous granite plutonism, i.e. contact and 'Barrovian' high temperature/low pressure metamorphism.

This contrast presents many interesting petrogenetic problems, especially when facies comparisons are attempted between Groix and Mesozoic/Tertiary blueschist belts of the circum-Pacific belt. Felix (1972) was the first to examine the facies status of the Groix glaucophane schists, and has suggested the possibility of an early glaucophane + lawsonite paragenesis that has subsequently suffered 'retrogression'.

For over 150 years, the Ile de Groix has been a favourite collecting ground for mineralogists in search of exotic and coarsely developed metamorphic minerals. An historical survey of this previous work is given in Cogné et al. (1966) and Mäkinen (1967). Only in the last decade has geological research concentrated on the structural (Cogné et al. 1966) and petrographic/mineralogical aspects (Mäkinen and Howie, 1972) of the glaucophane schist metamorphism. However, no attempts have been made as yet to synthesise these two bodies of data into a consistent petrogenetic model.

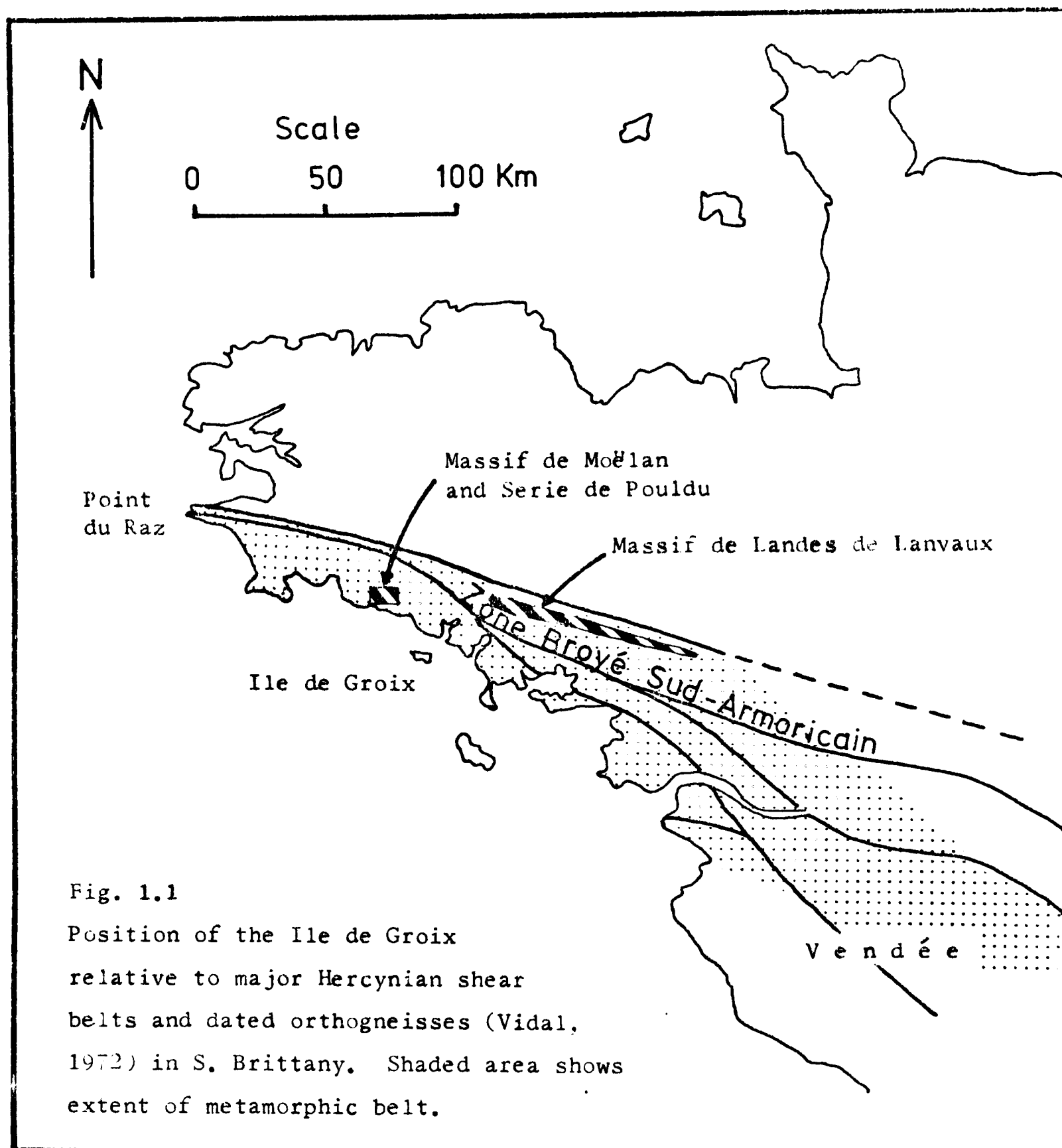


Fig. 1.1
Position of the Ile de Groix
relative to major Hercynian shear
belts and dated orthogneisses (Vidal,
1972) in S. Brittany. Shaded area shows
extent of metamorphic belt.

Triboulet (1974) has proposed a petrogenetic model based on mineral compositions in supposed equilibrium assemblages, but fails to take into account the textural relationships of the various analysed minerals to the different deformational episodes. Neither does he make use of the abundant source of chemical and mineralogical data in Makanjuola (1967) and Makanjuola and Howie (1972) to test out his hypothesis of metamorphic facies zonation on Groix. No reference, either, is made in any of Triboulet's work to Felix's structural-mineralogic study (1972) of lawsonite pseudomorphs on Groix, which is crucial for a fuller understanding of the facies-sequence as preserved in the metamorphic tectonites of Groix.

1.2. Aims of the present study

In the present work, it is proposed to use the available structural and mineralogical data of previous workers, where these data have been corroborated by observations of the present author, in the assessment of petrogenetic hypotheses.

In addition, structural and petrographic information is compiled from outcrops of differing bulk chemistry and metamorphic grade in an attempt to recognise a sequence of deformational events that may prove useful in unravelling the time sequence of mineral growth during metamorphism (see Chapter 3).

The main body of the petrogenetic argument is presented in Chapter 4 of the present work, where original microprobe analyses are used in conjunction with published experimental data and theoretical considerations to infer the conditions of metamorphism.

Having tackled the question: what kind of metamorphism?, the present author has proceeded with the next question: when was the metamorphism? (see Chapter 5). Very few extraneous assumptions are necessary for the interpretation of the original geochronological data, but those that

are made rely on the equilibrium temperature estimates of Chapter 4 and the trace element analyses given in Appendix II.

Chapter 6 is mainly concerned with the question How?, and the various regional implications of the petrogenetic model proposed by the present author are examined. Also, further tests and checks on the given hypotheses are suggested.

Throughout the discussion of petrogenetic interpretations of the chemical and mineralogical data a definitive basis of rock nomenclature is used and a clear division of the rock types is effected into chemical and/or paragenetic groups - this classification is provided in Chapter 2.

1.3. Outline of regional geology

According to Cogné (1960) and Cogné et al., (1966) the Ile de Groix rocks have suffered a long polymetamorphic history, in some ways comparable to the history inferred from supposed basement/cover relationships in the Anticlinal de Cornouaille (see Fig. 1.2) on the mainland. Groix is thought to lie on the southern margin of this polymetamorphic belt, at such a tectonic level that the more complexly folded 'basement' is seen beneath a more simply folded, younger 'cover'. The basement 'infrastructure' is assumed to have developed in the Cadomian orogeny (690-570 Myr ago, C.J. Adams, 1967) during amphibolite grade metamorphism of Brioverian sediments (pre 690 Myr). Later, during the Hercynian orogeny, deformation in the 'superstructure' was characterised by an open style of folding and a pervasive greenschist metamorphism was overprinted on pre-existing assemblages. On this model, glaucophane schist facies metamorphism is anomalous to both orogenic cycles, and is 'explained' by special interference effects in the zone where 'infrastructure' meets 'superstructure' (Cogné et al., 1966, Ruhlmann & Jeannette, 1966).

The northern margin of the South Brittany metamorphic belt is marked

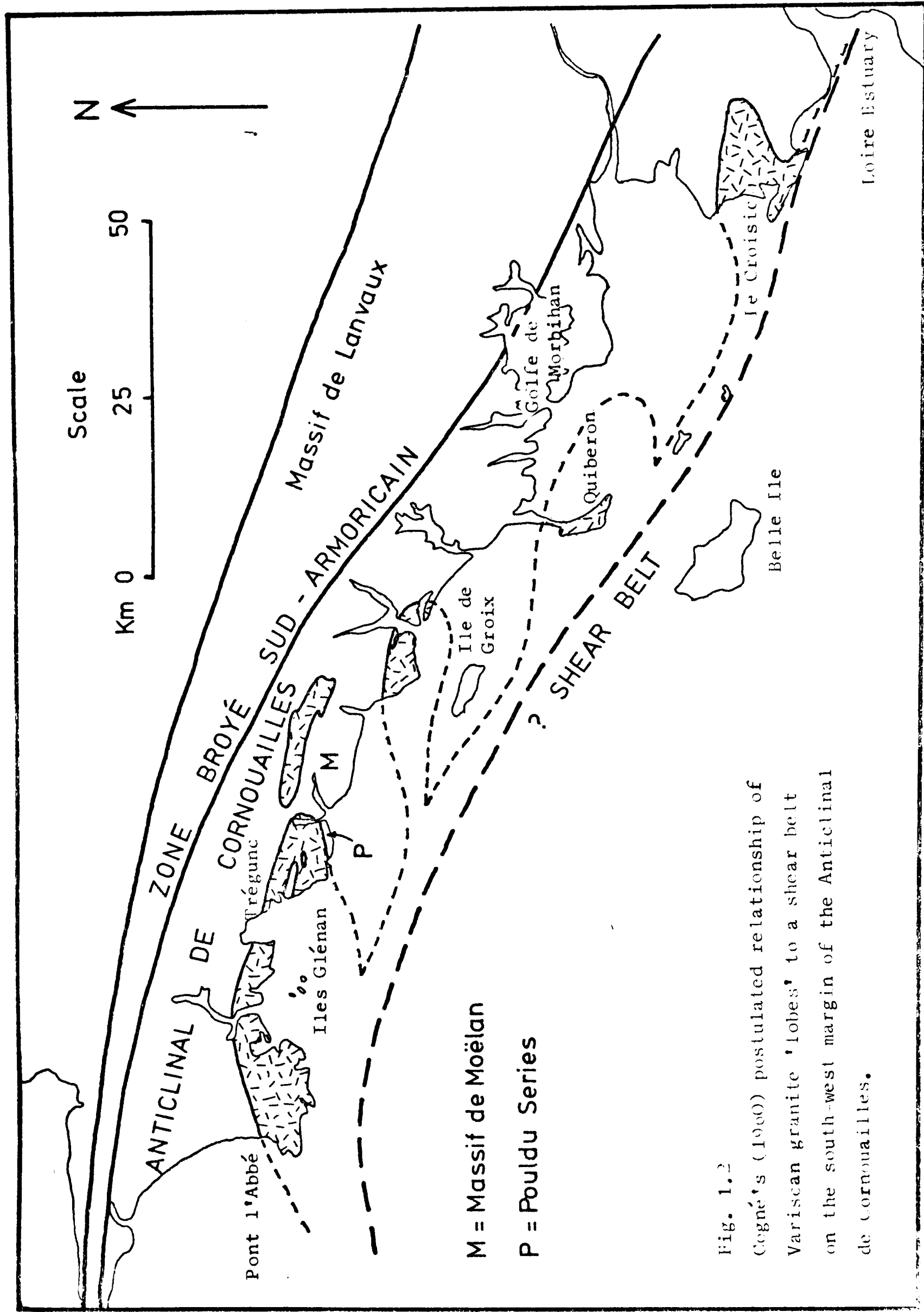


Fig. 1.2

Cogné's (1960) postulated relationship of Variscan granite 'lobes' to a shear belt on the south-west margin of the Anticlinal de cornoailles.

by the contact of highly sheared gneisses (such as the orthogneiss de Lanvaux) against almost unmetamorphosed sediments of Late Precambrian (Brioverian) or Lower Palaeozoic age that outcrop in an E-W trending syncline across Central Brittany (see Fig. 1.2).

The metamorphic belt itself (called the Anticlinal de Cornouailles in Finisterre and Morbihan by Cogné, 1960) is traversed by a series of important shear belts that cut indiscriminately across micaschists, granites and high grade gneisses and which appear to control the outcrop pattern of Variscan granites that intrude the surrounding rocks. The most important of these shear belts is the 'zone broyée sud-américaine' - a narrow (0.5 km) sub vertical belt of mylonites running for over 200 km with syn-tectonic 2-mica Variscan granites occurring as thin sheets or lobes along its length. Occasionally, these granite lobes escape from the confines of the shear zone and intrude Brioverian and Palaeozoic strata to the north (up to as young as Dinantian in the Chateaulin basin).

Most of the Variscan 2-mica granites have been dated (Adams, 1967, Vidal, 1973) at 340 ± 10 Myr old, an age which corresponds with the major movement on the shear zone and a slight unconformity at the base of the Carboniferous in the Central Brittany syncline (so-called Bretonic phase of the Hercynian orogeny, Cogné, 1960).

Within the gneiss core of the Anticlinal de Cornouailles, metamorphism reaches the sillimanite grade and migmatites are commonly observed. It still remains unclear whether these high grade rocks could represent an older 'Cadomian' basement, but the available geochronological data (Vidal, 1972, 1973) would suggest that the peak of metamorphism was reached in the Lower Palaeozoic (ca 450 Myr ago). The juxtaposition of the blueschists on the Ile de Groix with these high grade rocks on the mainland is at least suggestive of a paired metamorphic belt at some stage of the Palaeozoic in S. Brittany.

Another belt of Hercynian granites runs to the south of the gneiss

core' of the Anticlinal de Cornouailles, passing from Pont l'Abbé through the Iles Glénan to Quiberon and Le Croisic (see Fig. 1.2.). Most of this presumed granite belt is now underwater, and has been related by Cogné (1960) to an 'ad hoc' Hercynian shear belt, with associated intrusive lobes coming off to the north-east. If this speculation is correct, then the Ile de Groix is surrounded almost on all sides by Hercynian granites. However, no contact metamorphic effects are visible on Groix, unless one counts the sporadic occurrence of tourmaline as an indication of the proximity of granites.

1.4. General description of the Ile de Groix - fieldwork techniques

The Ile de Groix is a rocky plateau 8 km from north-west to south-east, generally less than 3 km broad, having the appearance of an outlier of the important post-Hercynian peneplane surface on the south Brittany mainland. Rising only about 40 m above sea level, this rock platform is bordered by steep, inaccessible cliffs on the windward side, which are cut by little clines or 'vallons'. A few more important valleys appear to be drowned, and form short, steep sided rias which afford good access to rock outcrops. On the leeward, south-east coast, cliffs are much lower and extensive wave cut platforms are developed (especially at the Pointe des Chats).

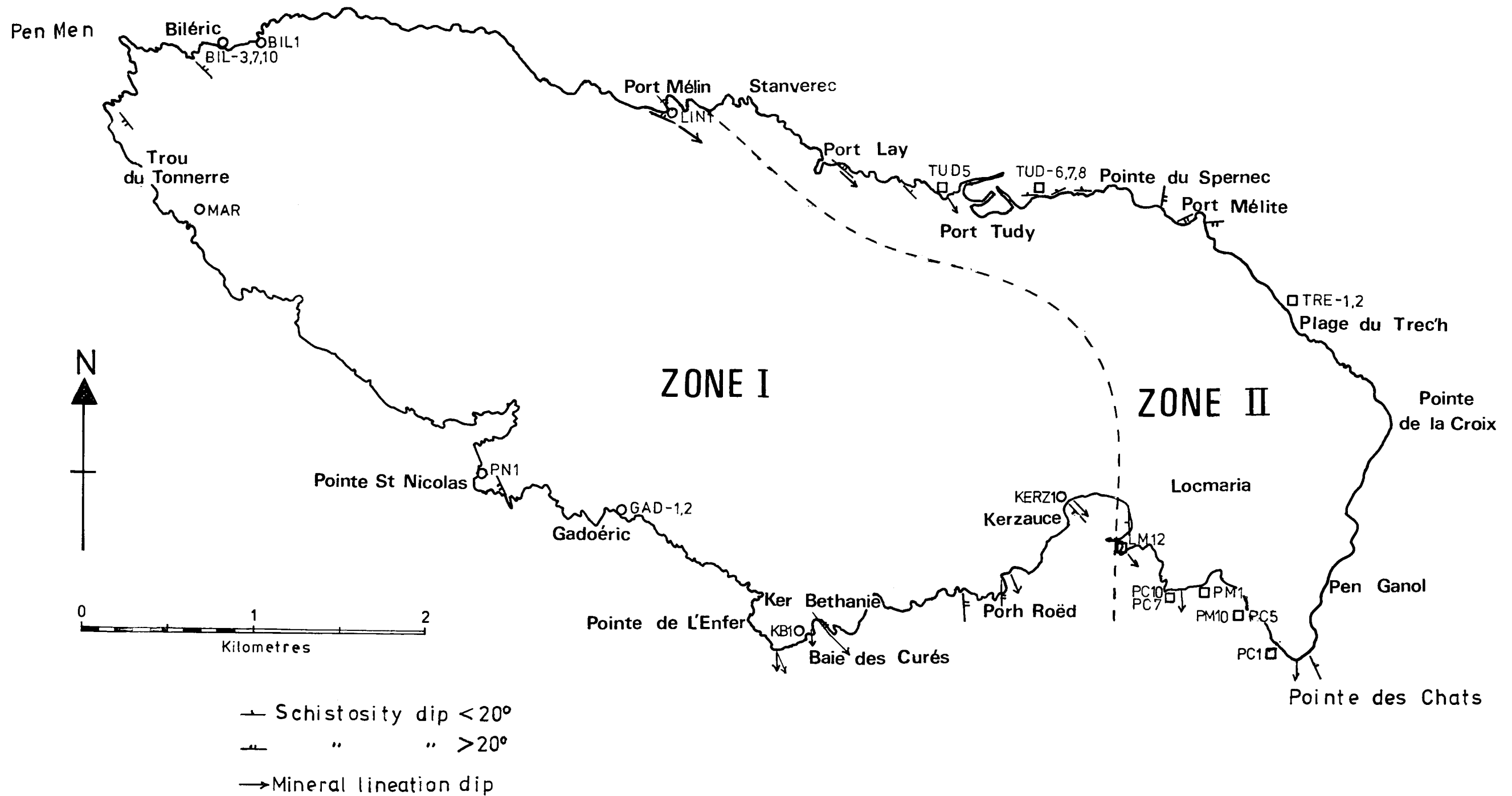
Rock outcrops on the Ile de Groix are almost exclusively restricted to the cliffs or fore-shore zones, thus making the compilation of a geological map very difficult, in view of the lack of inland exposure and the bewildering complexity of lithological interfingering on all scales.

Sample collecting for chemical and/or isotopic analysis was restricted to those outcrops of fresh appearance when broken. Wherever possible, samples were collected from outcrops above the HWM, so as to avoid the effects of recent sea-water contamination. The locations of

samples mentioned in the text of the present work are shown in Fig. 2.1, and the descriptions of analysed samples are listed in Appendix I.

Unfortunately, there are no accurate sample localities available for the mineral separates of Mäkanjuola (1967) used by the present author for geochronological purposes.

Fig 2.I Sample localities and distribution of prograde metamorphic zones



CHAPTER 2

CLASSIFICATION AND DESCRIPTION OF ROCK TYPES EXPOSED ON THE

ILE DE GROIX

Introduction

Makanjuola (1967) has proposed a system of dividing the rocks of Groix into chemical groups, using molecular proportions of the major element oxides CaO , Al_2O_3 , MgO in whole rock samples, and then subdividing these groups further on the basis of observed mineral paragenesis.

	<u>Rock groups</u>		<u>Rock type sub-divisions</u>
I	Aluminous schists	A	Glaucophane schists
II	Calcareous schists	B	Chloritoid schists
III	Magnesian schists	AB	Glaucophane + chloritoid schists
IV	Basic schists	C	Non-glaucophane, non chloritoid schists

Thus any rock may be classified, if the whole rock chemistry is known, by combining the appropriate epithets, e.g. IVC, a non-glaucophane non-chloritoid bearing basic schist.

Unfortunately, this system of classification neglects the difficulties of sample heterogeneity and hand specimen identification of crucial, yet sporadically developed minerals. Furthermore, a bulk chemical analysis is not always available to assist in the naming of a rock.

The present author has preferred a system of nomenclature that relies on easily recognisable features in outcrop or hand specimen, though it is still useful to retain such broad chemical groupings as 'aluminous' and 'basic'.

Aluminous schists are by far the commonest rock type on Groix, accounting for over 70% of the rock area at present exposed. The aluminous schists include such diverse lithologies as quartz-sericitic

phyllites, graphitic schists, chloritoid micaschists, albitic micaschists and garnet micaschists, and are characterised by low density (2.8 - 2.9), light colouration and a gross preponderance of white mica in the mode (> 36% by volume).

Basic schists of one sort or another, make up most of the remainder of the area of outcrop on Groix (25%) and ^{are} interfoliated with the aluminous schists on all scales from several cm to 20 m.

Of this 25%, may be 10% is made up of rocks rich in glaucophane (30% in mode) and the remaining 15% is accounted for by greenstones rich in chlorite and albite.

'Calcareous' (sensu Mäkinen, 1967) and magnesian schists are very restricted in occurrence and minor in volumetric importance. Magnesian schists are restricted exclusively to one outcrop in the Crique de Bilheric, whereas 'calcareous' schists are always associated with thin marginal zones around greenstone bodies or glaucophane schist bodies.

In this chapter, the rocks of the Ile de Groix are divided into four broad groupings:

- (1) Aluminous schists
- (2) Banded epidote schists
- (3) Magnesian schists
- (4) Massive basic schists

These groupings are then subdivided on the basis of observed mineral paragenesis on the hand specimen scale, and the definitive assemblage is listed at the head of every lithology description.

At the end of this chapter, information about the geographical distribution of mineral parageneses in each rock grouping is summarized in Table 4. Sample localities, where known, are indicated on Fig. 1 along with the division between prograde metamorphic zones (I and II) proposed by the present author.

TABLE 1

Average of modal analyses in % volume

Rock type	Phengite	Quartz	Glauc	Ctd	Alb	Garnet	Epidote	Chl	Rut/ilm	Sphene
IA Aluminous glaucophane schists (5 specimens)	52.65	23.51	7.09		(1.96)	(4.95)	(14.77)	(2.70)	4.8	
IB Aluminous chloritoid schists (10 specimens)	53.35	(59.16)		14.84		(6.76)		8.62	2.03	
IAB Aluminous glaucophane-chloritoid schists (3 specimens)	43.96	18.12	15.72	10.95		(2.90)	(0.96)		(0.93)	(5.57)
IC Aluminous non-glaucophane non-chloritoid schists										
a) Quartz-poor albite schists (3 specimens)	41.38	1.31			28.96		(16.25)	(21.72)	8.95	(0.58)
Quartz-rich albitic schists (5 specimens)	40.44	40.69			14.65	(1.50)	(0.95)	(5.22)	(0.68)	(4.12)
b) Quartzose micaschists (3 specimens)	36.51	54.79				(8.67)	(4.60)	(3.0)	0.32	

All data extracted from Mäkinen (1967), Chapter II. Modal analyses in brackets are for non-essential phases and are maximum measured values. Abbreviations are Glauc. for glaucophane, Ctd. for chloritoid, Alb. for albite, Chl. for chlorite and Rut/ilrn for Rutile and/or ilmenite.

(1) The Aluminous Schists

It should be stressed that the following description of the aluminous schists of Groix is based on:

- a) Mäkinen and Howie's (1972) and Triboulet's (1974) published mineral analyses from these schists.
- b) Micrometric modal analyses reported by Mäkinen in his thesis (1967).
- c) A comparison of the present author's thin section and field observations with those of Mäkinen (1967).

The following treatment of the above material is not intended to be exhaustive, but rather to serve as a guide to identification and classification. Salient petrographic points, especially regarding metamorphic grade and bulk chemical control on mineral paragenesis and composition, will be mentioned within each sub-heading of rock type and discussed in more detail in Chapters 3 and 4. A further aim is to de-limit the geographical extent of various metamorphic facies in rocks of constant composition. At the head of every section, a definitive mineral assemblage is given, in order of decreasing modal abundance for the rocks concerned. Non-essential phases (that is, phases which may or may not occur according to grade or minor chemical variation) are indicated by a $\pm$ symbol.

For the classification of the aluminous schists, the system used by Mäkinen (1967) is retained, apart from a further subdivision of Group IC to accommodate quartz-poor albitic schists.

IA Aluminous glaucophane schists

Phengite + quartz + glaucophane $\pm$ epidote $\pm$ garnet $\pm$ chlorite

Mäkinen (1967) analysed 5 schists of this type, of which two, from Kermarec and Locqueltas in Zone I (see Fig. 1) are non-garnet bearing. This geographical distribution suggests that the presence of garnet in

this lithology is grade controlled.

Garnet in this common pelitic lithology occurs as small, inclusion free euhedra (0.3 mm diameter) that are undeformed and devoid of reaction rims. (An analysis of a garnet, LM8, from this lithology, quoted by Mäkinen and Howie (1972), gives 10% spessartine, 69% almandine, 6% grossular, 10% andradite and 5% pyrope). These garnets very much resemble the small euhedral garnets described by the present author in some metabasic rocks (present study: PHOTO Fig. 3.12) and probably record the same episode of mineral growth (post F_1). More rarely, the garnets form large (5 mm) prophyroblasts with swirls of rutile and epidote inclusions (Mäkinen, 1967, Fig 2.5).

The blue amphiboles, which are an important accessory mineral in these schists (17% max mode), have never been analysed chemically, but their optical properties are closely similar to the glaucophane-ferroglaucophane series. These blue amphiboles are commonly pseudomorphed by chlorite, albite, magnetite and quartz to produce rhomb shaped bodies (Mäkinen, 1967, Fig. 2.3 and c.f. spec 97862.1 of Harker Collection Cambridge). According to Mäkinen, the glaucophane growth is occasionally pre-tectonic to a strong F_1 foliation, and augen-like structures are formed. On the other hand, cases have been seen by Mäkinen (1967, Fig. 2.4) where glaucophane crystallization is clearly post-tectonic, overgrowing minute garnet euhedra and producing a fish-net inclusion pattern with respect to quartz.

Epidote occurs, along with ilmenite, as helicitic inclusion trails in large garnets, and itself includes rutile and ilmenite. From this Mäkinen (op.cit.) deduces that, in the aluminous glaucophane schists, epidote began crystallizing before garnet and glaucophane but later than rutile and ilmenite.

It is important to differentiate the different habits of crystall-

ization of the white micas in these schists, since they appear to have developed at different stages of the deformational history, and show pronounced geographical gradations. One habit, termed sericitic by Mäkinen (op.cit.) is characterized by strongly micro-folded, fine grained white mica, and is typical of phyllites from the low grade, non garnet-bearing facies in the west of the island (Zone I). (The other habit is of elongate, unstrained, non-polygonized laths up to several cm long which appear texturally younger, and is typical of the higher grade schists in the west of the island.

Sometimes the two habits can occur together in the same rock, and it is not known whether there are present two micas of contrasting composition. Mäkinen and Howie (1972) have established that the 0.1 - 0.3 mm size fraction mica is of uniform phengite ($\text{Si}^{4+} = 6.52$ per 22 (O)) composition for high grade aluminous schists.

IB Aluminous chloritoid schists

Phengite + chloritoid + chlorite $\pm$ quartz $\pm$ garnet

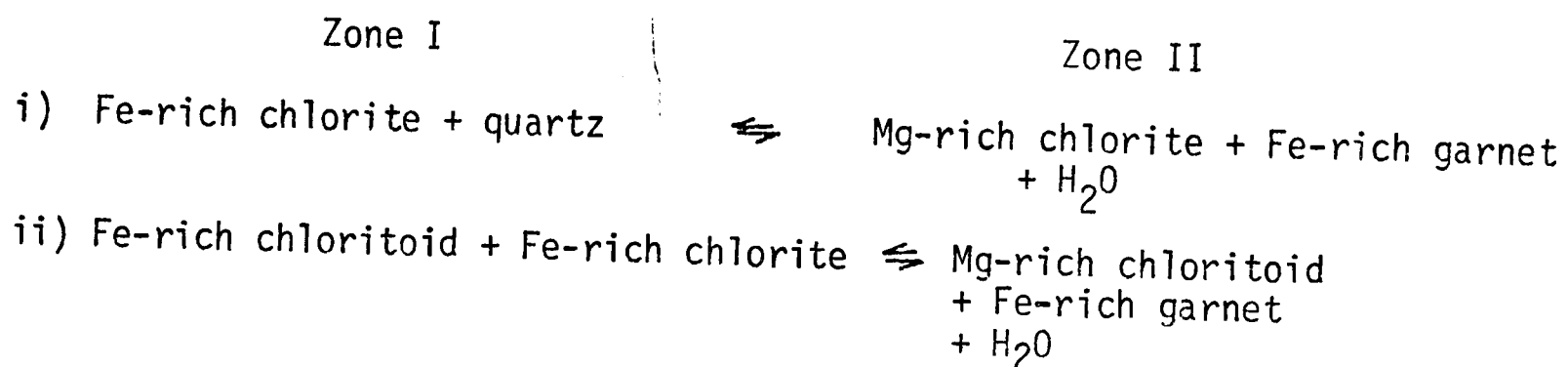
Of the 10 analysed schists of this type quoted by Mäkinen (1967), 6 are garnet-bearing varieties from between Locmaria and the Pointe des Chats. At Port Mélice, on the N.E. coast, there is an outcrop of aluminous schist with 5 mm porphyroblasts of chloritoid (present author's spec MEL1). The lithology exposed here is distinctive in having an intense, parallel and planar schistosity, S_1 , with isolated quartz rods formed from boudinaged isoclinal fold cores. The chloritoids are only faintly lineated.

Amongst the aluminous chloritoid schists, two broad textural groups can be recognised; fine grained (phyllitic) and coarse grained (porphyroblastic) schists. From Mäkinen's data (op.cit.) alone it is difficult to deduce the precise distribution of these schist types, but again it appears that the coarsely porphyroblastic schists

belong to Zone II (see Fig. 1) and the phyllites, generally lacking garnet, belong to the lower grade Zone I.

The phyllites show a strong continuous banding of chloritoid and quartzo-sericitic layers, which is often disrupted by a vertical strain slip cleavage with associated recrystallization.

Chloritoids from the Zone I non garnet-bearing phyllites (e.g. MK68 of Mäkinen and Howie, 1972) are noticeably weaker in their pleochroism, and have very much higher Fe^{2+}/Mg ratios (e.g. 7.8 in MK68) than the chloritoids from aluminous schists in Zone II (average Fe^{2+}/Mg in Zone II chloritoids = 3.0, Mäkinen and Howie, 1972). The higher Fe^{2+}/Mg ratio of chloritoids from non garnet-bearing aluminous schists is very likely a function of grade rather than whole rock composition since, in the absence of other important ferromagnesian minerals, the chloritoid and chlorite might be expected to accommodate the available Fe. i.e.:-



In the coarsely porphyroblastic chloritoid schists, the Mg-rich strongly pleochroic chloritoids are seen to cut across the fine grained 'sericitic' or parallel felted white mica, which is often microfolded, and forms embayed grain boundaries (Mäkinen, 1967, Fig. 2.8). It appears that the garnet and chloritoid are in equilibrium but post-date the 'sericitic' white mica, which contains inclusions only of rutile and ilmenite. The porphyroblastic laths of mica, on the other hand, contain inclusions of garnet, chloritoid and quartz as well as rutile, and thus are thought to have crystallized after the 'sericitic' mica, and post-tectonic to F_1 as recorded in inclusion trails.

Compositionally, the garnets (LM5, LH5 of Mäkanjuola and Howie 1972) are almost identical with those garnets analysed by the same authors in the glaucophane-bearing aluminous schists (e.g. LM8), but are larger, more subhedral and altered, with zoned arrangements of quartz inclusions. Mäkanjuola (1967) has interpreted the elongation of these garnets parallel to schistosity as evidence for their deformation during growth, but the present author sees no reason why slightly elongate garnets could not be the result of post-tectonic mimetic crystallization. Alteration cracks in the garnets could well be due to late retrogressive events rather than being related to early deformation of the garnets.

IAB Aluminous glaucophane-chloritoid schists

Phengite + glaucophane + quartz + chloritoid + chlorite
 ± garnet ± epidote ± sphene

This is rather a rare paragenesis, represented in Mäkanjuola's (1967) sampling by only four specimens, three of which were analysed. The present author has located only one outcrop, corresponding with Mäkanjuola's (op.cit.) PM6 locality near Stang er Marc'h, where this unusual lithology is developed in association with highly sheared quartz mica schists.

These aluminous glaucophane + chloritoid schists are distinct from all other types of chloritoid + glaucophane rocks on Groix (such as described by Kienast and Triboulet, 1972, and those in group IIAB of Mäkanjuola, 1967, and the present author's specimen PM10) by their great modal abundance of phengite (ca. 40%) and almost complete lack of epidote. In field relations they are distinct because they are not intimately associated with the margins of basic bodies, as are all other epidote-bearing glaucophane + chloritoid schists.

In common with the other aluminous schists, these schists have

rutile inclusion trails that appear to trace out an early schistosity (S_0) within the glaucophanes and chloritoids. But outside these coarse porphyroblasts (some chloritoids in PM6 are up to 14 mm long), and especially in contact with rare epidote, euhedral sphene is developed (forming 5% of the mode in PM6). Sphene may have grown at the expense of rutile and some early formed Ca-rich phase, such as lawsonite.



Lawsonite could have been stabilized in these schists during prograde metamorphism by an unusually low fO_2 - this is suggested by the very low Fe_2O_3/FeO ratio (0.035) of Makanjuola's PM6 whole rock sample. Also, the present author has observed rectangular and irregular patches of albite and white mica with scattered laths of epidote in sharp contact with sphene euhedra (specimen MAR1 of present author). These pseudomorph patches lie in the micaceous bands between coarse porphyroblastic, zoned glaucophane and chloritoid.

Inclusion free, but cracked garnet euhedra are restricted to sporadic layers, and partly enclose epidote crystals. Thus, they seem to have grown at a late stage of the prograde metamorphism after sphene formation but before the observed breakdown of glaucophane to chlorite and magnetite.

IC Aluminous non-glaucophane, non-chloritoid schists

a) Albitic micaschists

(or 'feldspathized potassium-white mica schists' of Makanjuola, 1967. 'Gneisses albitiques' of Triboulet, 1974)

Phengite + albite + quartz $\pm$ chlorite $\pm$ epidote $\pm$ garnet $\pm$ sphene $\pm$ chloritoid

Chemically, a very heterogeneous group of light coloured, low density (2.7 - 2.8) schists that all contain albite (An3 optically

determined in specimen PT1 of present author). No chemical data are available for plagioclases from these schists, though the general lack of multiple twinning, high positive 2V and low relief with respect to quartz all suggest Anorthite contents < 3 mole %.

The composition of the white mica in these schists is known only in four samples analysed by Triboulet (1974) (PdGr2, CdS2, Amu 4, Sp2), and the Si^{4+} content per 22 (O) falls in the same narrow range (6.4 - 6.6) as the phengites from all other aluminous schists from Groix (Velde, 1967a).

The one chlorite analysed from this group (Triboulet, 1974; Poulz 2) gives the highest Fe^{2+}/Mg ratio and Al^{IV} substitution for Si^{4+} observed in any chlorite from Groix.

The albitic micaschists can be subdivided into two chemically and petrographically distinct sub-types not recognised by Makanjuola (1967), but distinguishable in the field.

One sub-type, which includes Makanjuola's specimens LSM1, PM1 and KTM9, is very low in modal quartz ($< 2.0\%$) but unusually rich in rutile/ilmenite (up to 12% in mode). These quartz-poor albitic micaschists contain predominantly 'gefüllte' albites; that is, albites full of inclusions arranged in trails parallel to early schistosity. The only mineral not to be included in these large, anhedral poeciloblasts, is chlorite which is confined generally to the 'sericitic' groundmass.

The other type of albitic micaschist is very rich in quartz (up to 60% in mode in Makanjuola's KM3) and poor in rutile/ilmenite. Sometimes, as in the present author's specimen PT1, the albites are polysynthetically twinned, inclusion poor and occur in close association with strained quartz grains of about the same grain size. Genetically speaking, the albites that are seen in the aluminous schists of Groix may represent two separate generations that have developed in rocks of contrasting origins at different times. Since rocks of very

dissimilar parentage may end up with very similar bulk chemistries and mineralogies (e.g. arkosic sandstones and granites; lamprophyres and analcime bearing clays and tuffs) the chemical and paragenetic way of classifying the Groix schists probably obscures any genetic relationships suggested by field and textural studies.

Thus, there are included in the sub-type of quartz-poor albitic micaschists various rocks of presumed igneous parentage (e.g. Makanjuola's PM1; present author's BIL7 and EF1) that containgefülte albite which has been slightly strained and elongated parallel to the foliation. In other rocks of this sub-type, of the same bulk chemistry, the albite occurs as twinned, inclusion-poor grains that may be unrecrystallized detrital relicts (e.g. Makanjuola's LSM1).

If these inclusion-poor, twinned albites really are relict detrital grains, then the parentage of this rock was probably sedimentary.

In the quartz-rich albitic micaschists, bothgefülte and 'detrital' albites are observed, and this can be straightforwardly interpreted as the result of metamorphism of acid igneous rocks and arkosic sandstones respectively.

The close field association of some albitic micaschists (e.g. present author's TUD6, PT2) with the margins of metabasic bodies is very marked, and may indicate an original pre-metamorphic association of rock types or syn-metamorphism soda metasomatism. (In the former case, acid and basic rocks could have been juxtaposed in the manner of a composite dyke, or local weathering of a basic lava flow could have given rise to analcime or albite rich clays).

Alternatively, a process of syn-metamorphic metasomatism may have operated very locally where P_{Fluids} was high and concentration gradients were steep. In any case, Na-enrichment in micaschists marginal to chlorite + albite greenstones is restricted to a zone no more than 2 m wide (e.g. present author's TUD6 - TUD7), so the process, whatever its origin, is volumetrically unimportant.

Other albite-rich schists (e.g. present author's BIL7, EF1 and Makanjuola's 'PM1', and a rock from Trou de Tonnère) are not marginal to metabasites bodies, but instead they have slightly cross-cutting relationships with the regional foliation (L. Richards pers. comm., 1975). Such an association of lithologies and field relations can be interpreted as a relict intrusive relationship between felsites or keratophyres and argillaceous country rocks. The chemical evidence (present author's analyses of BIL7, Makanjuola's analyses of PM1, KTM9) does not conflict with this hypothesis.

b) Quartzose micaschists

('schistes banales, schistes quartzo-sericiteuses' of Cogné 1960.

Quartz, potassium-white mica schists of Makanjuola, 1967).

Quartz + white mica + rutile ± garnet ± epidote ± chlorite ± magnetite
± graphite ± ? biotite

These are rocks of normal pelitic or semi-pelitic composition that have not developed chloritoid due to high Fe/Al ratios and generally low Al_2O_3 contents. They are, judging from the four chemical analyses of Makanjuola (1967), richer in SiO_2 than all the other aluminous schists apart from the quartz-rich albitic schists (SiO_2 range 62-72.6 wt.%), thus meriting the name quartzose micaschists.

The white mica in these schists, generally 'sericitic' in appearance is assumed to be phengite on the basis of six partial analyses quoted by Makanjuola (1967) from the rocks LM7, LSM11 and LM11. Otherwise no chemical information is available on the minerals from these schists apart from the partial, semi-quantitative microprobe analyses of Triboulet (1971), which show the presence of stilpnomelane and biotite.

In outcrop, these quartzose micaschists are characterised by abundant quartz veining parallel to schistosity, which is often folded and boudinaged by early deformations. Sometimes, as in the Baie des

Curés, graphite is copiously developed and this gives rise to a peculiar deformational style described as 'fuseaux' by Cogné et al. (1966) (see Fig. 3.4 of this study).

The major control on garnet development in these schists appears to be metamorphic grade, since all garnet bearing graphitic and quartzose micaschists are confined to Zone II.

Triboulet (1971 and 1974) has recorded the occurrence of brown biotite in equilibrium with garnet and phengite in albite and quartzosericitic schists from zones extending from the Pointe des Chats to Locmaria and from Fort Surville to Pointe de Spernec. Further west, quartzose micaschists are deemed (Triboulet, 1974) to have the assemblage:

Phengite + stilpnomelane + garnet + chlorite

where the tie-line biotite-garnet is not stable. At lower grades still, Triboulet maintains, the garnet-chlorite tie-line is unstable, and normal pelitic compositions in the west of the island can form the assemblage:

Phengite + chloritoid + stilpnomelane + chlorite

Unfortunately, this attempt to delineate sub-facies using pelitic assemblages is totally unsupported by the necessary whole rock or zone mineral data. At the highest grade, one might expect to see Fe-rich biotite replacing stilpnomelane within the garnet zone, as long as the whole rock composition were Al deficient:

Phengite + stilpnomelane + chlorite $\rightleftharpoons$ Fe-biotite + garnet
+ H₂O

The evidence for the brown pleochroic mineral observed in these high grade quartzose micaschists being iron-rich biotite rests on incomplete microprobe analyses and XRD determinations of (001) and (002) d spacings (Triboulet, 1971).

Elsewhere on the island, stilpnomelane appears to be scattered

sporadically amongst the pelitic and semi-pelitic schists, both as a primary mineral in equilibrium with chlorite or chloritoid and as a product of late retrogression in association with green biotite (Triboulet, 1971). Again, the evidence for the occurrence of stilpnomelane is based on partial microprobe analysis and characteristic d spacings. There is a possibility that many of the pleochroic micas recorded in the Groix rocks could be altered chlorite (chlorite-vermiculite) finely intergrown with phengite and chlorite. The optical properties of chlorite-vermiculite are very similar to those of biotite, and these two minerals have often been easily confused, as in the Otago schists of New Zealand (E.H. Brown, 1967).

(2) The Banded Epidote Schists

Introduction

It is in this section that the present author proposed a major deviation from Mäkinen's (1967) chemical grouping of rocks on Groix. The reason for this change is that the present author has detected a major overlap in bulk chemical composition between the so-called calcareous schists and basic schists of Mäkinen (Groups II and IV respectively) and therefore prefers to adhere rigidly to a paragenetic way of classifying the metabasics, coupled with a due appraisal of their degree of mineral segregation on the outcrop scale. This latter feature is not necessarily related to the paragenesis, but it is a great aid in quick field identification.

In this group of rocks, very many more data from the present author's observations are brought to bear on the problem of classification, hence there is generally less reliance on Mäkinen's modal tables and analyses. Modal averages for the essential minerals of each paragenetic type are thus considered inappropriate (in any case they are very difficult to assess realistically due to the heterogeneities).

The classification used here, based so firmly on observed paragenetic associations, is constantly cross-referred to a NA - C - FM - A chemographic tetrahedron, where the bulk chemistries and relevant constituent mineral compositions are projected from albite onto the A - C - FM plane. With the use of this projection (Fig. 2) the equilibrium phase assemblage for most 'basic' compositions can be predicted and compared with the observed paragenesis in the field.

Chemical and paragenetic variation of the banded epidote schists
(includes 'calcareous glaucophane schists' of Mäkinen (1967))

It is unwise to label these schists as 'calcareous' since, firstly, their CaO wt.% whole rock contents are strongly influenced by mm and cm scale monomineralic bands, and secondly, several so-called 'calcareous schists' overlap in their chemistry with rocks defined as 'basic glaucophane schists' (e.g. LH13 of Mäkinen (1967) has only 7.97 wt.% CaO, but is classified by him as a calcareous schist. Average of 15 basic schists gives CaO = 8.06 wt.%; average of 6 calcareous schists gives CaO = 12.55 wt.%).

These epidote rich (20 - 80% epidote in mode) rocks nearly always seem to occur near the sheared margins of basic glaucophane schist boudins or greenstone bodies, to which they probably bear some genetic relationship.

As well as containing tightly folded mm scale epidote layers these schists may display a bewildering variety of contrasting mineral assemblages in adjacent bands. Almost always, however, epidote is the dominant mineral, and although glaucophane has been recorded in all seven 'calcareous' schists analysed by Mäkinen (1967), there is no reason why non-glaucophane bearing garnet or chloritoid rich rocks should not be included in the general association of banded epidote schists. To be more precise, the only feature which distinguishes

the banded epidote schists paragenetically from some basic rocks is the absence of pyroxene. Otherwise, all the 'calcareous' schists of Mankajuola are identical in paragenesis to many basic schists of the present author (e.g. TUD5, PC5).

All the epidote banded schists analysed by the present author (apart from the chloritoid bearing PM10) are consistently higher in Al_2O_3 and CaO content than the associated pyroxene-bearing metabasics at the same grade. Thus, they tend to lie on the epidote-chlorite tie-line in an A - C - FM triangle (Fig. 2) in such a place that the assemblage glaucophane + lawsonite could also encompass the composition. Stable assemblages for this composition are as shown in Table 2.

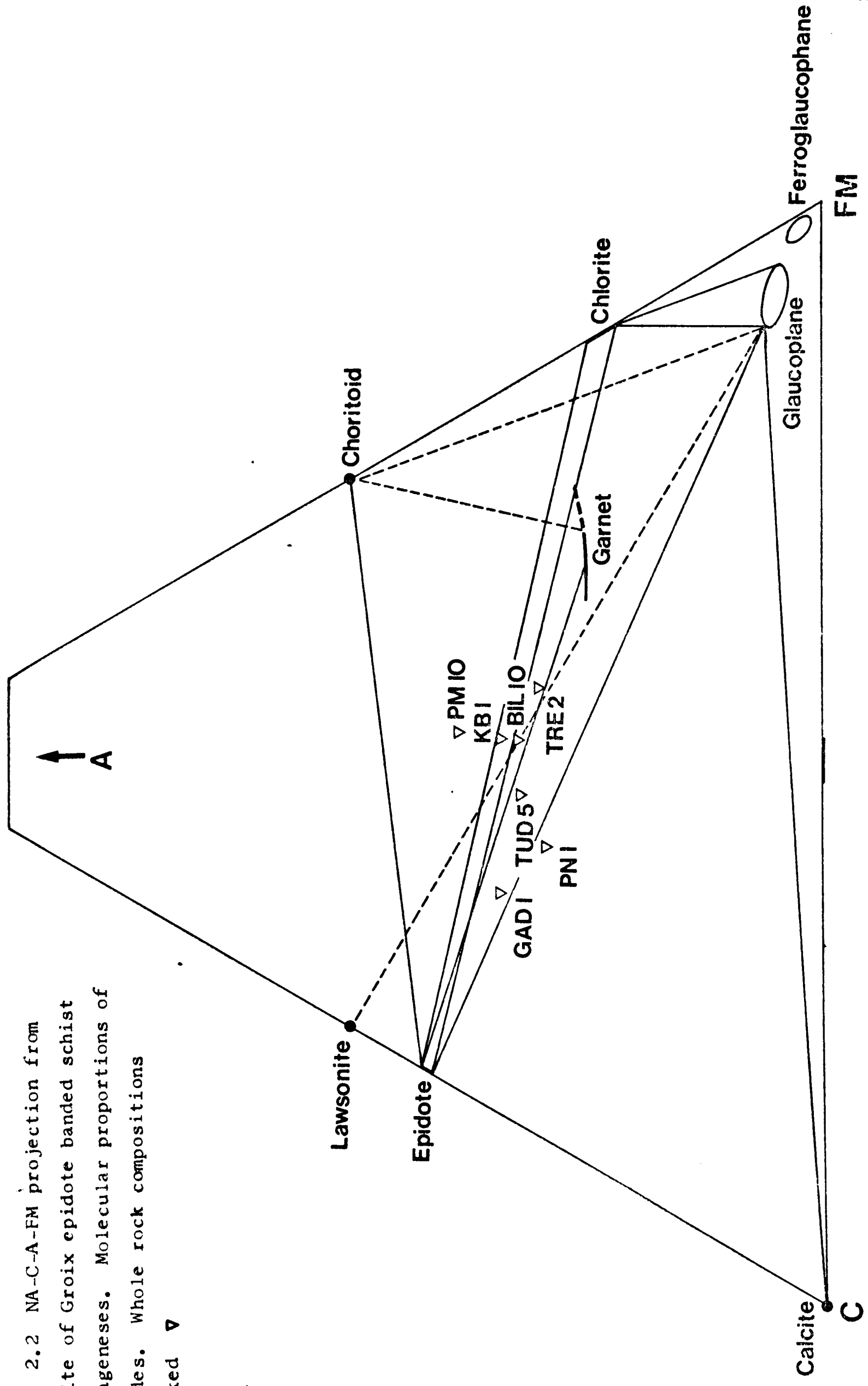
TABLE 2

LOW GRADE	HIGH GRADE
Epidote + chlorite + albite	Epidote + garnet + albite
Epidote + chlorite + glaucophane	Epidote + glaucophane + garnet
Epidote + calcite + glaucophane	Epidote + garnet + calcite

This is ignoring, for the moment, the possibility of lawsonite-bearing assemblages, and varying composition only towards the CaO component.

Another important feature of the banded epidote schists is the common occurrence of pseudomorphs after lawsonite, especially in rocks with Fe_2O_3/FeO ratios < 1.0 (Felix, 1972). The possibility of a stable glaucophane + lawsonite paragenesis during some stage of the prograde metamorphic history of Groix has been discussed by Felix (1972), who concludes that low fO_2 could have stabilized lawsonite in rocks of suitable composition at low grade. At higher grades, lawsonite probably broke down, even if fO_2 was falling throughout

Fig. 2.2 NA-C-A-FM projection from albite of Groix epidote banded schist parageneses. Molecular proportions of oxides. Whole rock compositions marked ▽



progressive metamorphism (see Chapter 4).

The mode of occurrence, characteristics and mineral assemblage of these lawsonite pseudomorphs are described by Felix (1972) and discussed further in Chapter 3 of the present work.

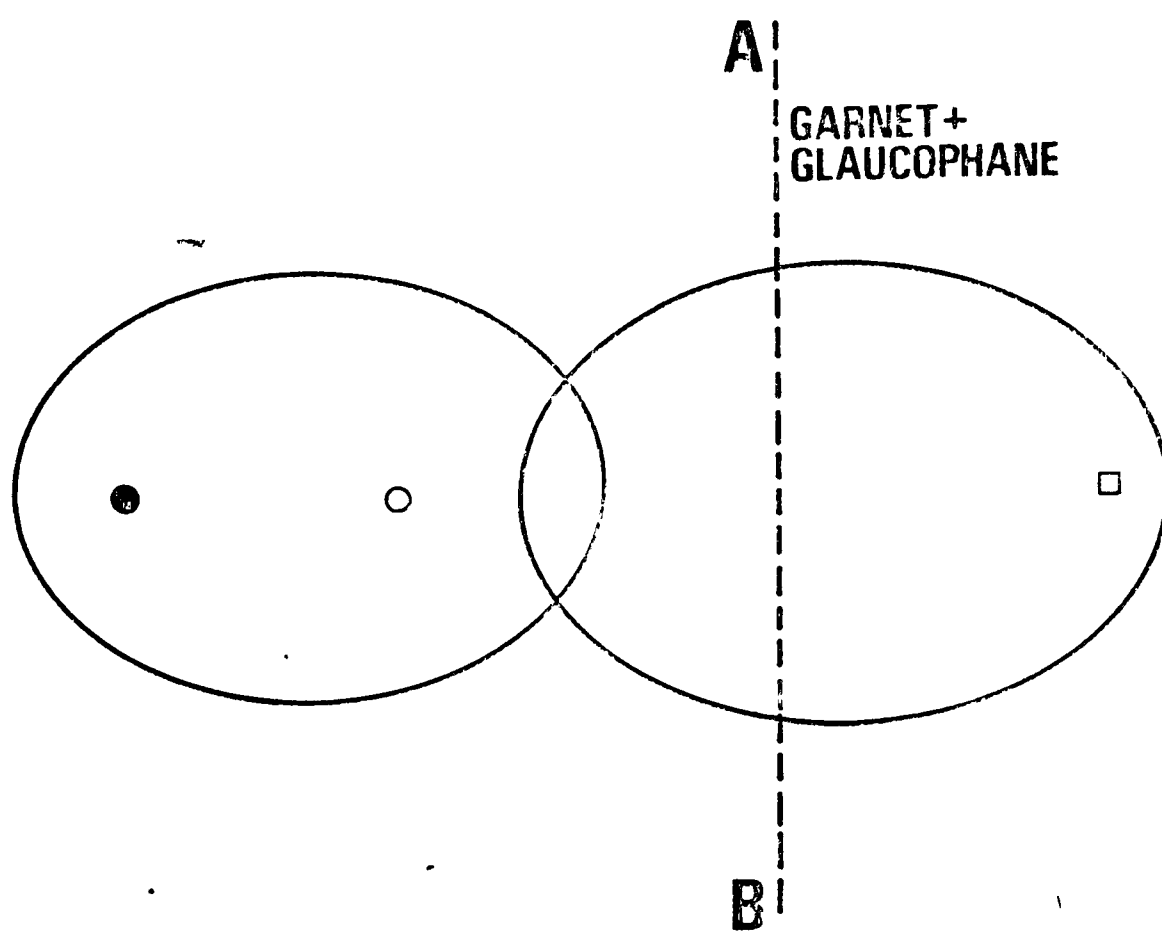
From the viewpoint of classification, all garnet-bearing 'calcareous' (Group II) schists of Mäkanjuola (1967) would be included in the present author's 'banded-epidote schists' along with the rocks described by Felix (1969) as 'glaucophanoschistes epidotiques à grènants'. In addition, various non-garnet bearing albite-rich rocks (e.g. present author's KBl, PMl), which contain the assemblage glaucophane + epidote + chlorite, can be regarded as the low grade isochemical equivalents of the garnet bearing types (see Table 2). The non-garnet bearing banded-epidote schists have been described by Felix (1969) as 'Prasinites à glaucophane et minéral noir', and are restricted geographically to Zone I.

In Fig. 3, the basic and calcareous schists of Mäkanjuola (1967) are visualized as two, slightly overlapping, compositional fields that develop sometimes common, sometimes distinct mineral assemblages according to grade.

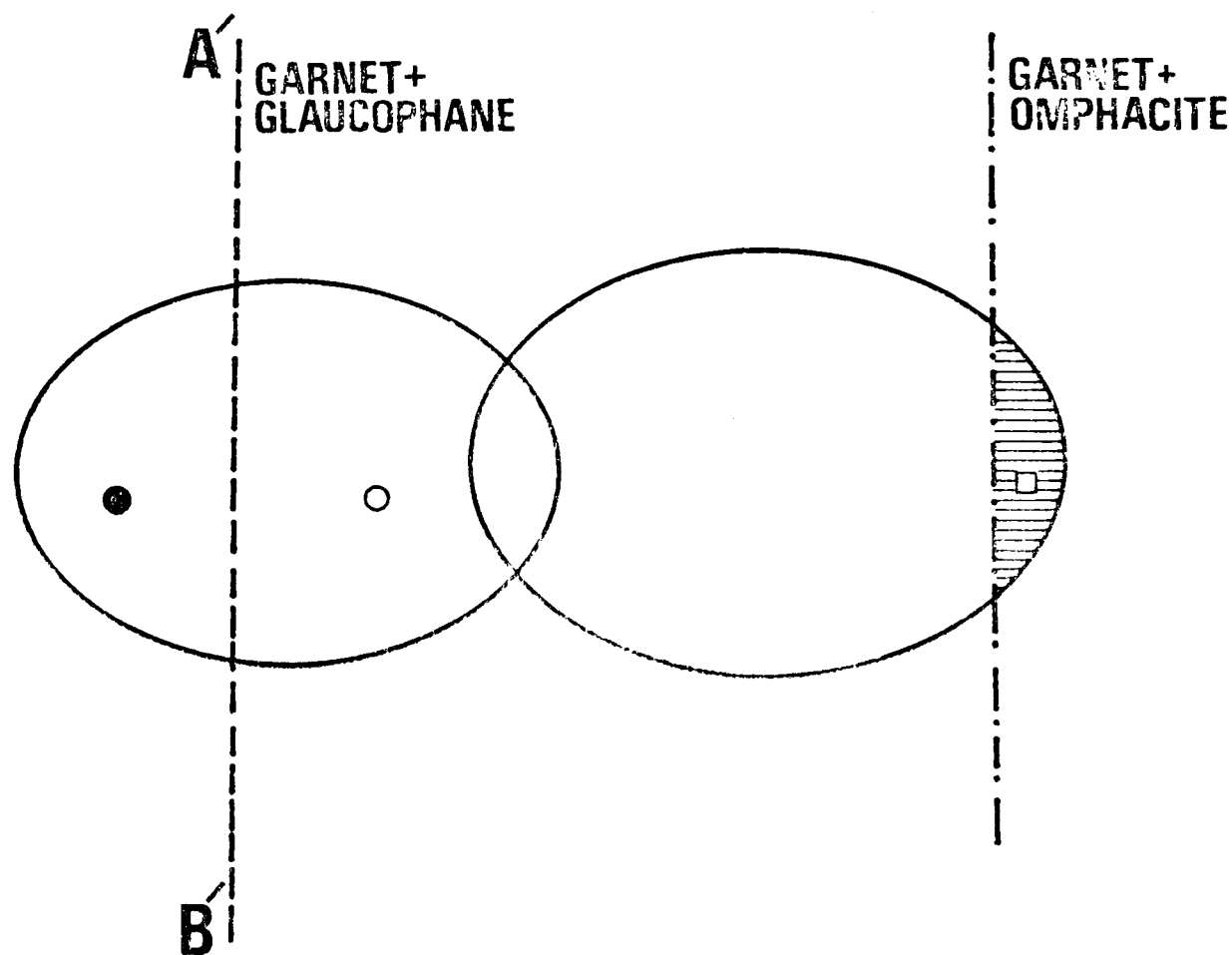
At low grade, only very Ca poor compositions lie to the right of a garnet forming isograd, A-B, whereas at higher grade, a garnet + glaucophane assemblage is common to a wide range of 'calcareous' and basic schists. Only the very Ca-poor rocks are able to form pyroxene bearing assemblages at high grade, and Ca rich rocks at the same grade may not develop garnet. With this in mind, it is possible to say that pyroxene bearing basic schists are always higher in grade than non-garnet bearing 'calcareous' schists regardless of composition, whereas garnet-bearing glaucophane schists could be isofacial with either extremes (though not both at once) depending on composition.

The present author's definition for 'epidote banded schists'

LOW GRADE



HIGH GRADE



←
Increasing CaO Content

Fig. 2.3 Compositional and grade control on banded epidote and massive basic schist parageneses.

thus excludes all pyroxene bearing basic schists, and includes many of the glaucophane + garnet + epidote basic schists (Type IV) of Mäkanjuola (1967).

In the majority of cases, the observed parageneses in the banded epidote schists correspond with the predicted equilibrium assemblages for rock compositions projecting (from albite) onto the epidote-garnet or chlorite join in A - C - FM space (see Fig. 2). Thus, it is thought that the system of classification proposed here reflects quite well natural, genetic groupings (according to grade, composition) and should be easy to apply in the field without prior knowledge of rock chemistry.

The present author recognises the following varieties, or common associations, amongst the 'epidote banded schists' of Groix.

TABLE 3

TYPE	Definitive assemblage	Mäkanjuola's (1967) samples	Present author's samples
a)	Epidote + glaucophane + garnet	Group II LH11, KM11, LH13 Group IVA LH4, LM4, 'LM12', LH14, LM15, LM17	TUD5, TRE2 QUED5, PM1, LAY1, GR1, SPE1
b)	Epidote + glaucophane + phengite	Group II LSM12	GAD1
c)	Epidote + glaucophane + chlorite + albite	Group IVA2, LM3, PM5 Group II, QM3, NH4	PN1, KB1
d)	Epidote + glaucophane + chloritoid + garnet	Group IIAB, PYM3 Group IVAB, LM16	PM10
e)	Epidote + garnet + chloritoid + albite		TUD 8, QUED 2
f)	Epidote + chlorite + albite		BIL10

a) Epidote + glaucophane + garnet $\pm$ phengite $\pm$ Qz $\pm$ chl $\pm$ Rutile

>> Sphene

This is probably the most common metabasic assemblage lacking albite or pyroxene to be found on Groix. Geographically, it is confined to areas east and north-east of the 'garnet in metabasics isograd' (Fig. 1) running through Locmaria and the Pointe de Stanverec. [Outcrops vary from a few centimetres to several metres in thickness, intercalated with aluminous schists and lying parallel to the dominant foliation. Banding within these schists is picked out by strongly microfolded, laterally persistent, epidote rich seams varying from less than 1 mm to 5 m in thickness. Between the epidote layers are sandwiched sometimes strongly lineated laths of glaucophane which may or may not show an orientation parallel to the B of observed isoclinal in the epidote layers. [The other essential mineral of these rocks, garnet, is either scattered as minute brownish euhedra (1 mm) as in spec PM1 of the present author, or can be found as larger 3-5 mm inclusion-rich porphyroblasts concentrated in layers and pods. In the latter case, inclusions of epidote, rutile and glaucophane are common and can be used as further criteria to differentiate these garnet bearing metabasics from other pyroxene bearing varieties.

Lawsonite pseudomorphs are quite commonly developed in these schists, especially in phengite rich layers, where they appear as rectangular or lozenge shaped prisms up to 2-3 cm long, often oriented in the same direction as the surrounding glaucophane. The pseudomorphs are composed of an aggregate of clinozoisite, paragonite and albite, the clinozoisite being concentrated in the core region of the pseudomorphs. Tiny, inclusion free euhedra of Mn-rich garnet (c.f. present author's PM1) are very often associated within the pseudomorphing assemblage.

Other characteristic features of these schists are the high density (3.2) and high modal abundance of rutile (up to 5%). Quartz

is nearly always present as an important accessory mineral, but its modal abundance in the rock is quite clearly related to the proximity of sheared contacts with aluminous schists (e.g. Mankanjuola's LH14 with 22.74% Qz in mode; present authors loca. GR1).

b) Epidote + glaucophane + phengite $\pm$ chl $\pm$ Qz

These schists are distinguished from type a) above by the absence of garnet and from type c) by the absence of albite. They are most distinctive in their type locality, Gadoeric, by their abrupt contact with massive chlorite + albite greenstones. [In appearance they are characterized by a rapid (mm scale) alternation of yellowish-green and dark grey/blue layers, the former being composed of epidote in very flattened and boudinaged isoclinal folds. The axes of these isoclinal folds are parallel to the contacts with encasing chlorite + albite greenstones. So this lithology can be considered as occupying a narrow (50 cm) shear zone between more competent, massive rocks on either side. The dark-grey blue layers in the rocks are made up mostly of glaucophane needles of variable size, the largest of which show a preferred lineation, within the planar fabric, parallel to the isoclinal axes (see PHOTO Fig 2.4).

Petrographically, the epidote occurs in two habits: firstly, as minute laths (2×0.05 mm) crystallized mimetically around micro-folds, and secondly as more equant grains in more quartz rich, coarsely crystallized layers. Glaucophane always shows varying stages of replacement by barroisite and/or chlorite and magnetite, and is often found as composite grains with barroisite rims.

This rock type has not been described previously on Groix, unless one counts the assemblage of rock 'LSM12' given by Mankanjuola, which bears a close similarity to GAD1 of the present author.

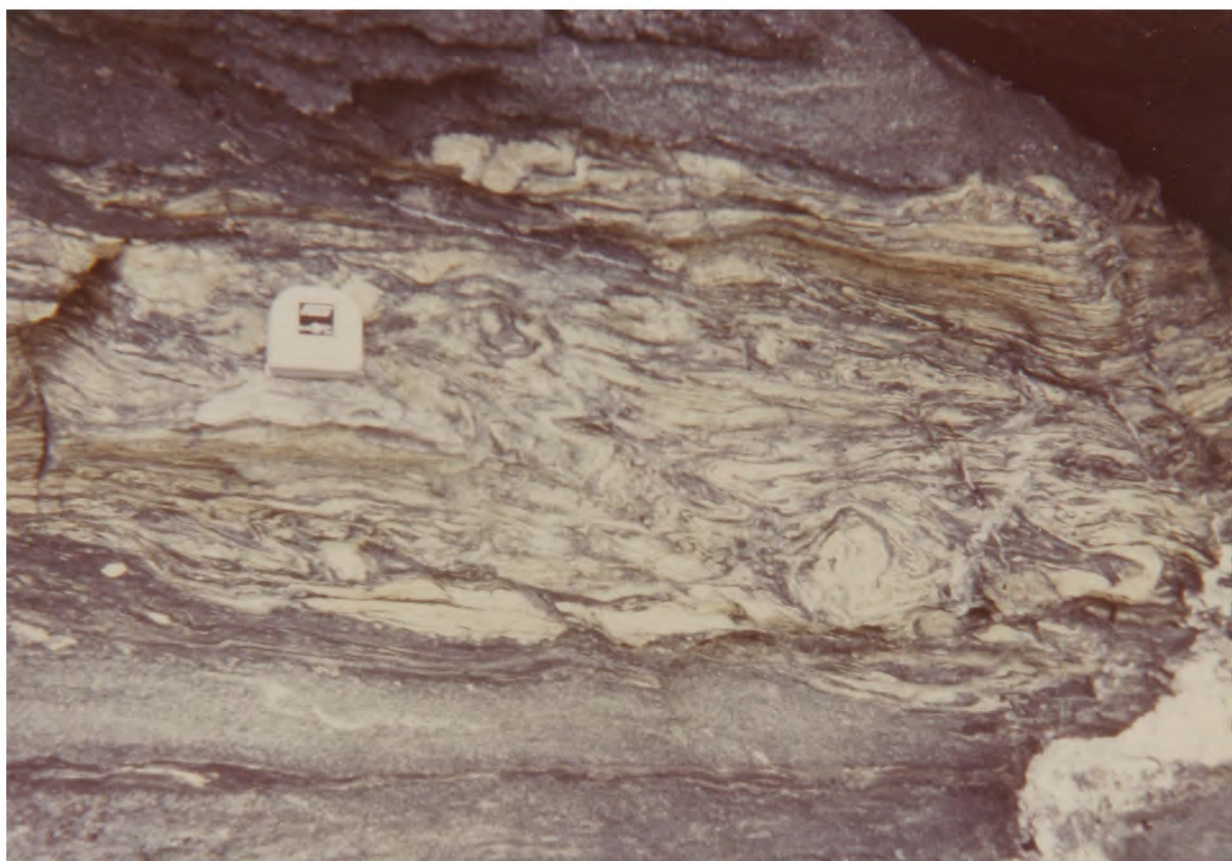


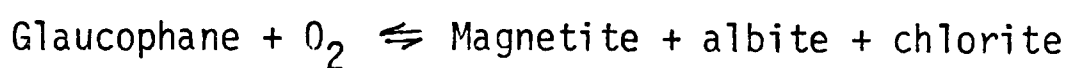
Fig. 2.4

Highly sheared, epidote banded schist with flat-lying F_1 isoclinal folds. Epidote rich bands are interbedded with glaucophane rich layers. Outcrop passes upwards directly into massive chlorite + albite greenstone (seen at top of photo). Locality GAD1 - Gadoéric. Scale is 5cm.

c) Epidote + glaucophane + chlorite + albite + calcite

These schists correspond in mineral assemblage to Felix's (1969) 'prasinities à glaucophane et minerais noirs' and to Mäkinen's (1967) non-garnet bearing calcareous schists and basic glaucophane schists, amphibole bearing sub-type. Although they vary in intensity of foliation and grain size, they can be recognised by the co-existence of glaucophane and albite in the absence of garnet.

In the more coarsely crystallized weakly banded varieties, glaucophane occurs as decussate laths up to 5 mm long, with a tendency to Fe-rich composition (e.g. present author's KBl). The laths are never bent, but show abundant signs of retrogression, along rims and cracks to either chlorite, albite and magnetite or barroisite. Albite occurs usually as small blebby inclusions within the glaucophane, or in an anhedral mosaic with quartz and white mica. [Epidote in these schists is almost invariably developed as small, stubby granoblastic grains rich in pistacite component ($P_s = 28-29\%$ in Mäkinen's NM4) and containing rutile inclusions. The epidote grains are often crystallized mimetically around microfolds, and this contrasts strongly with the coarse porphyroblastic epidotes from garnet bearing schists of this chemistry (type a). Magnetite is the 'minerais noirs' quoted by Felix (1969) as being so distinctive of this type of epidote banded schist, and is particularly abundant as octahedra near retrogressing glaucophane in the absence of garnet and omphacite. Viz:



Broadly speaking, there are two varieties of this type of epidote banded schist, one calcite bearing with or without sphene and the other calcite free with common lawsonite pseudomorphs.

i) Epidote + glaucophane + calcite + alb

ii) Epidote + glaucophane + chlorite + alb

The former is represented by the present author's specimen PN1, which corresponds in chemistry and paragenesis to Mäkanjuola's calcareous schist NM4. Both schists come from outcrops on the east side of Port St. Nicholas, and are interbedded with aluminous phyllites. Banding in this variety of schist is very marked, consisting of almost monomineralic epidote layers up to 1 cm thick folded into isoclinal folds. Brownish spots, composed of iron stained calcite, which are also characteristic of this rock type, probably do not represent pseudomorphs after some unknown mineral. Composite grains of sphene rimming rutile also seem characteristic of this schist and are rare in other types of epidote banded schist.

The other type of epidote + glaucophane + alb schist is calcite free, and is typified by the present author's KB1 specimen (similar to Mäkanjuola's QM3). Banding is much less marked than in the calcite bearing lithologies, and glaucophanes have crystallized as large crudely lineated or decussate laths. Magnetite euhedra are to be found in chlorite filled embayments and cracks in the glaucophanes, but no barroisitic rims occur. On weathered surfaces, orientated rectangular micaceous pseudomorphs, probably after lawsonite, are apparent. Sphene and apatite are important accessory minerals in these schists.

The type specimen for this variety of schist (KB1) comes from La Baie des Curées (see Fig. 1) near a blowhole on the west side. Here, the margin of the metabasic horizon is in contact with albitic quartz micaschists with small garnet relicts. The preferred orientation of the blue-black glaucophane laths gives a crude lineation within the foliation, perpendicular to both isoclinal fold axes and pseudomorph lineation ($B \rightarrow 130^\circ$).

It is likely that the type of glaucophane-albite epidote banded schist here described is the most common non-garnet bearing metabasic

assemblage on Groix after the massive chlorite + albite greenstones (see Fig. 3.13.).

d) Epidote + glauc + chloritoid + garnet $\pm$ alb $\pm$ paragonite $\pm$ phengite

This is a rare rock type on Groix which has nevertheless attracted particular attention due to the apparent stable co-existence of chloritoid and glaucophane, an unusual paragenesis restricted by special bulk chemical and pressure/temperature requirements (Hoschek, 1969). The present author has found only one outcrop, near Por'h Morvil between Locmaria and Pointe des Chats, that yields parageneses comparable with those described by Velde (1967b) and Kienast and Triboulet (1972). The bulk chemistry of the whole rock sample (PM10) is indistinguishable from most other basic schists (but quartz in the norm reaches 16%) and is closely similar to the rock LM16 analysed by Mäkinen.

In appearance, the rock is almost pegmatitic and strongly banded with non-orientated glaucophane euhedra up to 3 mm interlayered with isoclinally folded epidote rich layers, quartz veins and albite poeciloblasts. Where albite and white mica are abundant in the rock, glaucophane is absent and chloritoid appears to be strongly corroded. Garnets are scattered sporadically as large pinkish porphyroblasts, inclusion rich with chlorite reaction rims. At the locality of PM10, the described paragenesis occurs as a band at least 30 m long and 10-15 cm wide, sandwiched between a rather massive glaucophane + garnet + epidote horizon and a chloritoid micaschist.

There are no analyses available of minerals from PM10, but the compositions of co-existing glaucophane and chloritoid reported by Velde (1967b) from a rather more micaceous variety of the same paragenesis can be taken as representative of this group. On the basis of Velde's analyses, the chloritoid of this group is similar to all

the other strongly pleochroic chloritoids from garnet-bearing aluminous schists analysed by Mäkinen (1967), and has 25% substitution of Mg for Fe²⁺. This contrasts with the weakly pleochroic chloritoid (PM6) from the lower grade, non-garnet bearing aluminous chloritoid glaucophane schists which approaches the Fe-end member in composition. Glaucophane from this schist appears to differ from most other glaucophanes in garnet bearing metabasites by having high CaO content (3.65 wt.%), low 2V (-28°) and blue-green pleochroism. This analytical result may in part be due to a small admixture of barroisitic hornblende in Velde's (1967b) mineral separate (CaO = 7.71 wt.%, in Mäkinen LH14 barroisite).

Another important mineralogical feature of this rock type is the occurrence of paragonite, which rims corroded chloritoid near adjacent glaucophane grains. One can never observe straight unaltered contacts between glaucophane and chloritoid, since white mica and chlorite always intervene (see Figs. 14, Chapter 3). The white mica is assumed to contain some paragonite from the partial analysis quoted by Mäkinen (Na₂O = 5.54 wt.%, K₂O = 3%) for a white mica from his LM16 specimen.

There appears to be quite a lot of confusion as to whether the chloritoid forms an integral part of the assemblage with glaucophane, epidote and garnet. Mäkinen suggests that chloritoid has crystallized in this rock from late veins. But all the vein chloritoids from Groix have much lower Mg contents than the chloritoid that appears in this assemblage, and one would expect a stable Mg-rich chloritoid + garnet + epidote tie triangle for a rock of this composition at high grade. Fe-chloritoid would be stable with the low temperature assemblage and Mg-chloritoid, the one analysed by Velde (1967b) would be stable with the high temperature assemblages. From these considerations, one would expect to see lawsonite pseudomorphs

in this schist and evidence for the prograde breakdown of glaucophane. Both phenomena are observed as predicted by this model.

Phengite occurs as an important constituent of Velde's chloritoid + glaucophane rock, but the white mica in Makanjuola's rock seems to be predominantly paragonite, and makes up only 4.5% of the mode. Velde's (1967) chemical analysis of the phengite 'Br. Is.' shows it to be closely comparable in composition to the phengites from the aluminous schists analysed by Makanjuola. ($\text{Si}^{4+} = 6.65$ per 22 O).

e) Epidote + garnet + chloritoid + albite $\pm$ phengite

This paragenesis is found only in two samples collected by the present author (TUD8 and QUED 2) and is distinguished from d) by the absence of glaucophane. The rock type displaying this paragenesis is fine grained (0.1 - 0.2 mm) but has a few highly altered garnet porphyroblasts up to 2 mm in diameter. Very fine grained granular epidote is the dominant constituent of the groundmass, with some sericitic mica and chlorite. Chloritoid occurs as turquoise-colourless pleochroic laths from 0.25 - 0.5 mm long, often as gefüllte type inclusions in albite poeciloblasts (0.5 mm patches), in micaceous areas of the rock. These albite poeciloblasts are in optical continuity, untwinned and have lozenge or rectangular outlines similar to lawsonite pseudomorphs seen elsewhere on Groix.

In outcrop, this rock type is restricted to thin bands adjacent to massive basic bodies, and is characterised by a tough splintery fracture and mauve to slaty blue weathering colour. Lawsonite pseudomorphs up to 0.5 cm long are very apparent on wave polished surfaces, and show a strong preferred orientation parallel with local fold axes.

No whole rock or mineral analyses are available for this rock type, and no previous record of its occurrence is to be found. Nevertheless, it is possible on paragenetic grounds alone to classify it

with the 'banded epidote schists', probably being very similar in composition to PM10. The assemblage Epidote + garnet + chloritoid + albite would develop as a stable paragenesis at high grade after all glaucophane and lawsonite had broken down. Both of the samples collected by the present author come from outcrops on the eastern coast of Groix. TUD8 comes from Pen er Lan east of Port Tudy, and QUED2 from the east end of the Plate de Port Quedoul.

f) Epidote + chlorite + albite $\pm$ ilm + sphene

This is a common paragenesis among the low grade basic rocks of Groix which have either never developed glaucophane, or have lost it through retrogression. The paragenesis as quoted above would include all the basic non-glaucophane, non-chloritoid schists of Makanjuola (type IVC) as well as more epidote-rich, banded schists recognised and analysed by the present author (e.g. BIL10, BIL3).

It is only this latter group which, by virtue of their fine-grained, schistose texture and abundance of epidote, can be usefully classified with the 'banded epidote schists'. All the other basic non-glaucophane, non-chloritoid schists described by Makanjuola have epidote only as an accessory phase ($\leq 10\%$ in mode) and their texture is dominated by large (5 mm)gefüllte poeciloblasts of albite that cut across and obscure any schistosity the rock might have. By contrast, the epidote-rich banded schists of this paragenesis are much finer grained, granoblastic epidote is concentrated into almost monomineralic layers and 2-3 mm augen and the albite poeciloblasts have grown mimetically to the foliation. Parallel to this foliation, but never included in the epidotes, are abundant elongate grains of ilmenite, a feature which is rarely seen in the massive chlorite + albite greenstones, where opaque minerals are poorly developed.

At their type locality in the Crique de Bilêric on the north-west

coast of Groix, these epidote-rich chlorite + albite rocks have a compact, finely banded texture which contrasts markedly with the cavernous weathering, coarse, massive chlorite + albite greenstones with which they are in direct contact (see PHOTO Fig. 6).

It is proposed that glaucophane $\pm$ lawsonite was never a part of the primary, prograde metamorphic assemblage here, even if the bulk chemistry of the rock was favourable, since high fO_2 during metamorphism inhibited the growth of lawsonite, and excess Na_2O . Al_2O_3 was taken up in albite. This proposition is supported by the whole rock composition of the epidote + chlorite + albite rock BIL10, which could be encompassed by the assemblages glaucophane + lawsonite or chlorite + epidote + albite. Since no glaucophane relicts or lawsonite pseudomorphs are seen in this rock, and an early schistosity is preserved that is obliterated in massive chlorite + albite greenstones, then it follows that chlorite + epidote + albite was the stable equilibrium paragenesis in BIL10 before H_2O rich recrystallization in adjacent rocks caused the growth of largegefüllte albites and clinozoisites so typical of 'massive' greenstones. The only visible effects of this retrogressive metamorphic event on BIL10 would be the addition of magnetite octahedra, which are seen to include, and thus post date, the sphenes. The assumption about high fO_2 during prograde metamorphism of this rock is in part borne out by the high Fe_2O_3/FeO ratio (1.52) compared with massive greenstones.

In the field BIL10 is seen to grade more or less abruptly into a highly retrogressed greenstone about 1 m wide which is in contact with an H_2O -rich quartz-poor albitic aluminous schist (c.f. specimen BIL7, $H_2O = 5.5\%$). This latter lithology was probably a water donor during the retrogression, and one might expect intensity of retrogression to fall off with increasing distance from the donor. The field relations and structural-mineralogical features of the rocks in the outcrop are consistent with this 'retrogression control by

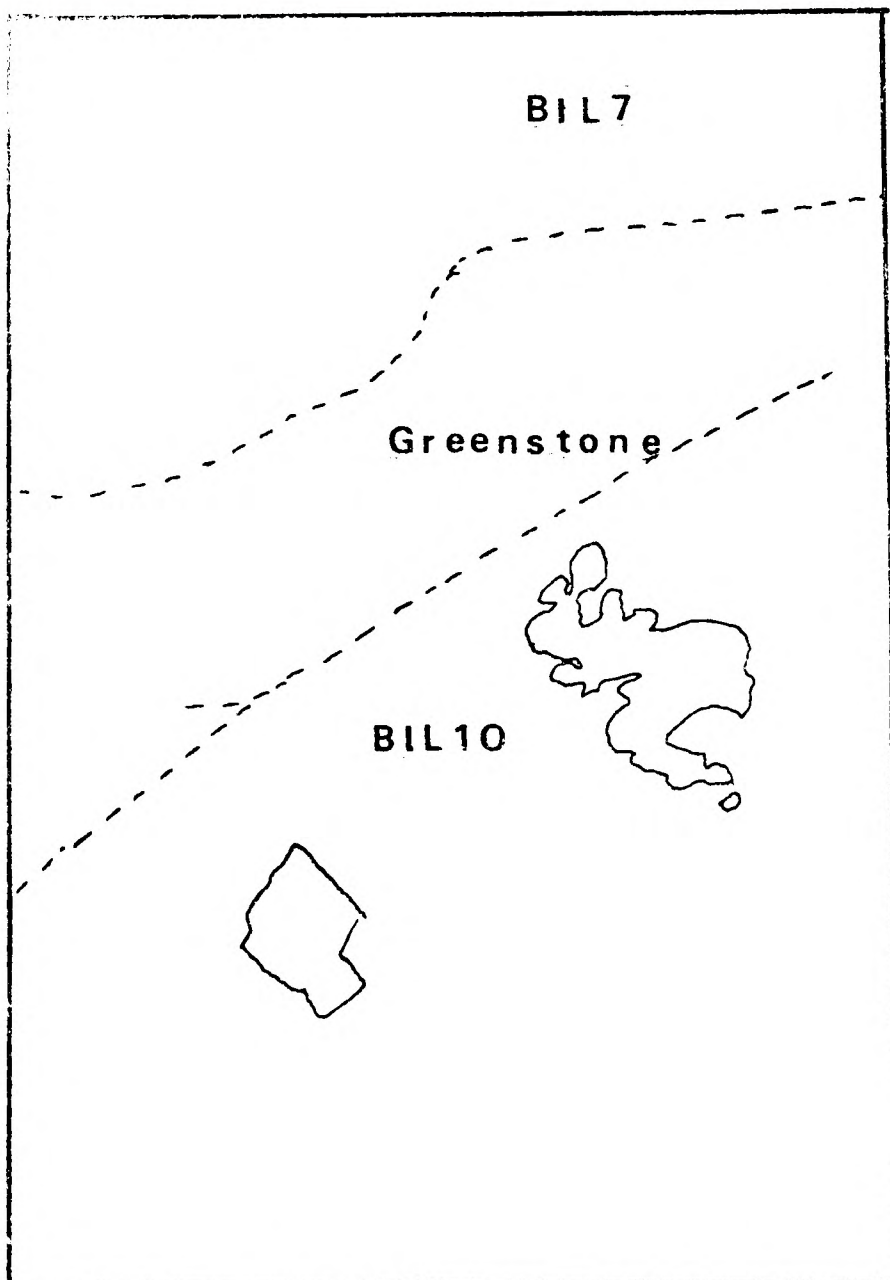


Fig. 2.6 Field relationships at the crique de Bileric locality. Hammer 30cm long. Explanation in text.

water availability' hypothesis.

(3) Magnesian schists

Talc + Actinolite + Chlorite $\pm$ mt $\pm$ Cr pheng

This is a chemically distinct group of rocks, first recognised and analysed by Mekanjuola, and restricted in outcrop on Groix to a small cliff ledge between the Crique de Bilheric and the semaphore house. Here, a massive talc + chlorite schist with magnetite octahedra up to 3 mm across, and radiating, decussate actinolites is in contact with a highly contorted chlorite + albite greenstone with albite poeciloblasts elongated parallel to the axial plane of the fold (PHOTO Fig. 7). In appearance, the rock varies from a soft fine grained talc-rich type with abundant magnetite octahedra to a more coarsely crystallized, dark brown weathering variety with radiating sheafs of actinolite needles, some up to 2.5 cm long.

It is very likely that this paragenesis grew very late in the metamorphic and tectonic history of Groix, since the actinolites are completely undeformed and non-orientated even in close proximity to greenstones which themselves suffered retrogression prior to a late north-east trending open fold phase (F_4) (see Fig 2.7).

Compositionally, the magnesian schists are unique amongst the rocks of Groix in having very high MgO contents (18 - 25 wt.%) low Al_2O_3 (3-4%) and appreciable concentrations of Cr_2O_3 (ca. 1.06%). This latter peculiarity is reflected in the composition of a mariposite (BM1) analysed by Mekanjuola which lies off the usual muscovite-celadonite solid solution exhibited by Groix micas. Although Mekanjuola did not analyse for Cr_2O_3 in this green mica from the magnesian schists, he inferred that it might well be a Cr bearing variety since the whole rock from which the phengite was separated contains 0.98 wt.% Cr_2O_3 .

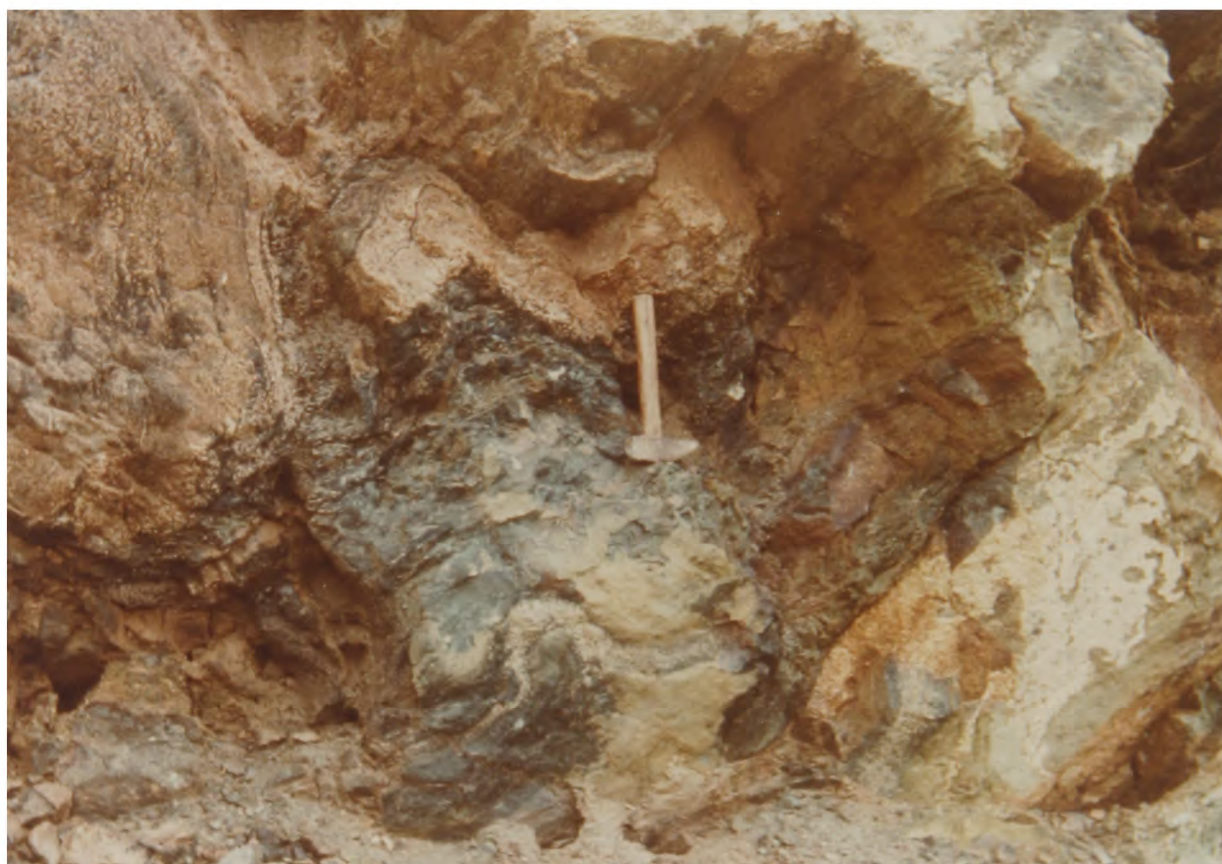


Fig. 2.7

F_4 fold in chlorite + albite greenstone band (top of photo). Albite euhedra are arranged parallel to axial surface of F_4 fold. Outcrop passes down into talc + actinolite + chlorite 'magnesian' schists with magnetite (dark green). Locality BIL1 - between Sémaphore and Biléric. Hammer shaft is 30cm long.

The magnesian schists can be thought of as originally ultra-basic bodies, that have been hydrothermally altered during a late retrogressive thermal event on Groix. It is unlikely that they represent rocks derived from CO_2 metasomatism associated with serpentinite bodies because no serpentinites have been discovered in situ anywhere on Groix.

(4) Massive basic schists

- a) Pyroxene-bearing glaucophane garnet schists
or glaucophanic eclogites
- b) Chlorite + albite greenstones
- c) Actinolite + albite + chlorite + epidote

Included in this group are all the basic rocks regardless of paragenesis, that will not fit under the broad classification of banded epidote schists due either to their lack of cm scale mineral layering, or low modal abundance (< 10%) of epidote or both. It follows that massive basic schists can exist with parageneses similar in most respects to those of the banded epidote schists already described, but remain distinct from the latter by their grain size, colour and texture. It is these last mentioned characteristics which enable a quick and easy identification in the field without recourse to thin section recognition of critical assemblages.

a) Glaucophanic eclogites or pyroxene-bearing glaucophane garnet schists

Glaucophane/barroisite + omphacite + garnet + epidote + sphene

± Quartz

This is a rock type of great petrogenetic importance, on account of its affinities with type C eclogites from the Franciscan blueschist terrain (Coleman et al., 1965). Unlike the Franciscan eclogites, which occur as exotic blocks within a structural mélange, these

glaucophanic eclogites from Groix are preserved in boudins that show a mineral facies and tectonic fabric concordant with the encasing micaschists. In addition, sphene is the stable Ti-bearing phase in these rocks as opposed to rutile, which characterizes the Franciscan type IV exotic blocks.

In the field, these rocks can be recognised by their dark colour (slate grey-yellowish green), high density (3.3) and fine grain size (< 0.3 mm). The finer grained varieties are poorer in pyroxene and richer in glaucophane and epidote, with small (0.3 mm) brownish garnets weathering out as spots. Although the rock is very tough, compact and homogeneous, a clear parallel alignment of the acicular glaucophane grains is apparent on close inspection. Often this mineral lineation is parallel to the long axis of the boudin within which the lithology is enveloped.

Coarser varieties are more yellowish-green in appearance, due to the development of omphacite laths up to 4 mm long and 1 mm wide in the 'pressure shadows' of garnet porphyroblasts. This habit leads to a streaky or augen like structure around garnets, which is often aligned parallel to the glaucophane lineation (PHOTO, Fig 2.8) of locality LM2. [As previously noted by Mäkinen (1967 thesis, Fig. 2.31) there is often a textural contrast in the same thin section between coarse, pyroxene-rich augen and adjacent fine grained glaucophane + epidote areas. Since sphene is the only included phase in the pyroxenes, and rutile is common as inclusions in the epidotes and glaucophanes, it is at least possible that pyroxene growth in these schists post-dated the breakdown of rutile in the epidote + glaucophane assemblage. Large garnet porphyroblasts (up to 3 mm) in these schists are seen to include sphene, epidote and pyroxene, so some or all of their growth must have overlapped with that of the augen. Straight, unaltered contacts are preserved between the



Fig. 2.8

Contact between massive glaucophanic eclogite and banded epidote schist horizon (30cm thick). Locality: Locmaria, near eastern jetty (LM12).

garnet rims and glaucophane and even small omphacite grains are included within the rims of the garnets, thus indicating that glaucophane and omphacite were both in equilibrium with garnet at the same time. But omphacite and glaucophane are always separated by a screen composed of barroisite/hornblende, showing the two minerals to be in reaction relationship during a late stage of the metamorphism. When one takes into account the frequent occurrence in some basic schists of composite rutile/sphene species with rutile cores and sphene rims, it then becomes easy to account for the development of pyroxene + sphene augen and sphene inclusions in garnet if one postulates a late, high temperature, low fH_2O stage of prograde metamorphism that has partially effected glaucophane + epidote + garnet metabasites according to availability of H_2O . Barroisite would have developed in these rocks wherever glaucophane came into contact with the dehydrated augen and whenever H_2O was locally available (see Chapter 3).

In terms of available mineral and whole rock analyses, these schists are amongst the most intensively studied rocks of Groix. Mäkinen (1967) has quoted 15 whole rock major element analyses of his 'pyroxene bearing sub-type of basic glaucophane schist'. Of these rocks, only eight actually contain omphacite in the mode, while the others are included whether or not they contain omphacite on the basis of having sphene inclusions in the garnets. He argues that the non-omphacite bearing glaucophane garnet schists with sphene inclusions belong to the same group as the pyroxene-bearing varieties since they both contain primary sphene, not rutile. This attempt at a sub-division of the basic glaucophane schists is fraught with difficulties, not the least of which is the impossibility of observing sphene inclusions in garnet porphyroblasts less than 3 mm across in the field.

In the absence of data on the type and intensity of mineral foliation in Makanjuola's (1967) specimens, it is not possible to decide whether some of his non-pyroxene bearing basic glaucophane schists should be described as banded epidote schists. However, it is safe to say that the higher the modal abundance of epidote in a rock, the lower is the likelihood of finding omphacite and the greater the intensity of metamorphic banding.

Compositionally, the massive pyroxene-bearing rocks of the present author (specimens PC10, PC7, LM12) are systematically lower in CaO and Al_2O_3 than the basic epidote banded schists, and thus tend to lie on the intersection of garnet-omphacite and glaucophane-epidote tielines in Na-A-C-FM space (see Fig. 9). Compatible mineral assemblages would be:

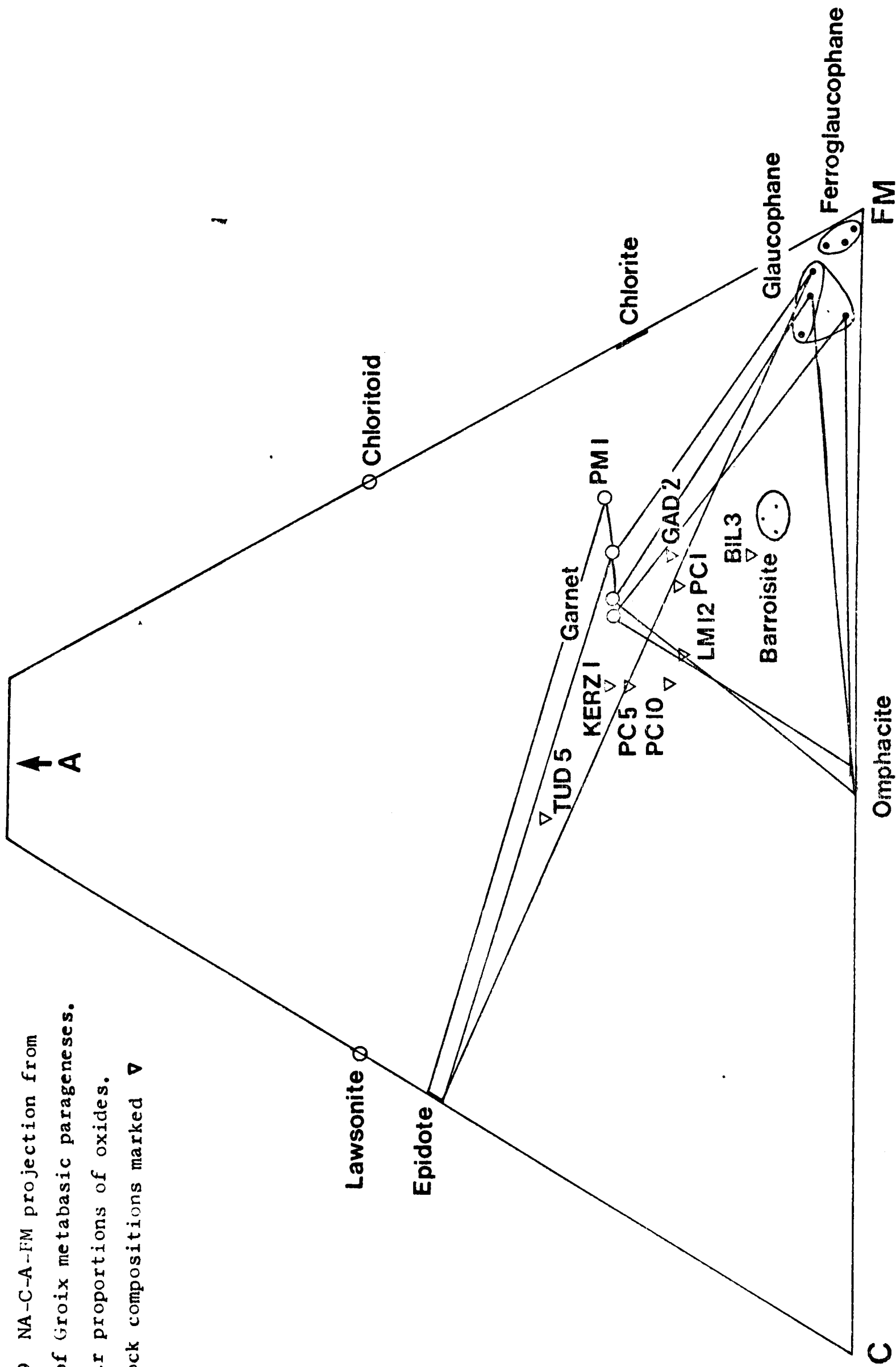
1. Glaucophane + Garnet + epidote + quartz : "WET"
2. Garnet + omphacite + albite + quartz : "DRY"
3. Epidote + omphacite + glaucophane + quartz

Assemblage 1 is commonly developed in domains surrounding the augen (Assemblage 2). Assemblage 3 is never observed, since glaucophane and omphacite never share common grain boundaries. It is Assemblage 2 which, in association with sphene and phengite, has provided so much useful information about the pressure/temperature conditions of prograde metamorphism on Groix (see Chapter 4).

Geographically speaking, the pyroxene bearing glaucophane schists collected by the present author all come from the stretch of coast between the jetty east of Locmaria, and the lighthouse at Pointe des Chats (specimens PC10, PC7 and LM12, see Fig. 1).

Other reported occurrences are the eight sample localities of Makanjuola (1967), seven of which are between Locmaria and Pointe des Chats, and one (PXMI) from the Côte d'Hène west of Port Tudy on the north coast. Triboulet (1974) attempts to de-limit a zone of highest

Fig. 2.9 NA-C-A-FM projection from albite of Groix metabasic parageneses. Molecular proportions of oxides. Whole rock compositions marked ▽



grade on Groix in which the assemblage 'glaucophane + garnet + omphacite + epidote' is 'very widespread' intercalated with garnet + chloritoid aluminous schists and Fe-biotite + garnet quartzose mica-schists. According to Triboulet (1974), this zone of highest grade is restricted to between Locmaria and Pointe des Chats on the south coast and between Pointe de Spernec and Fort Surville on the north-east coast. Such a zoneographic distribution of metamorphic facies on Groix fails to take into account the outcrops of pyroxene bearing metabasics west of Port Tudy. (Specimens mentioned by Brière 1920, from "la petite anse du Grippe voisine du bourg").

b) Chlorite + albite greenstones

Chlorite + albite + clinozoisite $\pm$ actinolite + sphene

These are rocks which are characterised by a massive, brownish cavernous weathering aspect, low density (2.83) and great modal predominance of chlorite (40%) and albite (30%) over any other constituent phases. Invariably, the albite has crystallized as large anhedral poeciloblasts (up to 5 mm across) that are untwinned and full of inclusions of clinozoisite, sphene and green amphibole. Chlorite is generally excluded from the albite poeciloblasts, and forms either crudely orientated wisps or large radiating clusters of lamellae (c.f. present author's specimens LIN-1, GAD2).

Makanjuola (1967) has observed both epidote sensu strictu and clinozoisite as inclusions in the gefülte albites. The clinozoisite is recognisable as first order grey birefringent, optically positive (high 2V) anhedral, whose low pistacite content (14%) is confirmed by microprobe analyses of the present author (c.f. LIN-1 Clinozoisite analysis). Epidote sensu strictu has never been analysed in these rocks, but it is assumed to occur as the higher birefringent, optically negative small grains, not necessarily confined to the albite poecil-

blasts, which also occur in some chlorite albite greenstones.

Slightly aluminous actinolites have been analysed in these schists by Triboulet (1974) (Krlv 3: amphibolite à chlorite) and by the present author (LIN-1: chlorite + albite greenstone). Triboulet says that, in common with alb + epidote amphibolites, the aluminous actinolites of the greenstones sometimes show a zoning towards a more aluminous rim composition, with blue green pleochroism (c.f. LH14 barroisite of Mekanjuola), and that discrete grains of both amphibole never occur together in the same rock. This indirectly supports the observations of Mekanjuola (1967) who states that ragged, ill-preserved relicts of optically negative pale yellow-pale green pleochroic amphibole (which he has misidentified as hornblende) are to be found, as well as more easily recognisable actinolite needles in the same group of rocks. This observation is best interpreted, in the absence of any positive proof for the occurrence of primary hornblende or barroisite in these rocks, as showing that aluminous actinolite—regardless of crystalline habit, is the only primary amphibole of these schists.

The present author has not in fact observed the zoning of actinolites towards more barroisitic compositions, and the analyses given by Triboulet (1974) refer to zoning only in amphiboles of the glaucophane bearing schists.

Another important mineralogical feature of these rocks is the apparent absence of rutile, either as tiny inclusions or as large porphyroblasts, and the abundant development of sphene. Elsewhere, sphene is seen as the mineral replacing rutile during prograde metamorphism or retrogression. Thus sphene would be the only Ti-bearing phase in both high grade cpx bearing blueschists and totally recrystallized, retrogressed chlorite + albite greenstones. Magnetite, although common as a breakdown product of partially retrogressed

glaucophanes, is very rare in the chlorite + albite greenstones due to the fact that retrogression has caused complete obliteration of the pre-existing assemblages and structures, with a consequent re-distribution of Fe.

In composition, the chlorite + albite greenstones are in no way different from most banded epidote schists and massive glaucophanic eclogites, except for their H_2O content, which averages slightly higher than other non garnet-bearing metabasics but ranges up to a maximum which exceeds any other rock type on Groix. (Mean H_2O = 3.4%, Max = 6-7%, Felix 1972, Triboulet 1974). This extreme state of hydration is also reflected in the generally high values of Fe_2O_3/FeO recorded for these rocks (1 - 3.5).

Since chlorite + albite greenstone bodies are distributed throughout Groix without any regard to tectonic style or prograde metamorphic zones, it seems reasonable to conclude their formation was connected with some retrogressive thermal event that post-dated prograde mineral growth. There is very strong evidence in outcrop, where a gradation can be seen between foliated glaucophane + garnet + epidote schists and massive chlorite + albite rock with relict garnets (PHOTO Fig. 10, locality PC1), to support an hypothesis of late, post-deformational, isochemical, water-rich retrogression.

The present author has collected chlorite + albite greenstones from horizons on Groix that vary in thickness from 20 cm (TUD7) to 20 m (Locqueltas). Generally, the massive greenstones are interbedded with aluminous or albitic schists that show some signs of local metasomatism, (fall off in mode and grain size of albite away from contact: development of monomineralic epidote layers in basic schists, i.e. GAD1).

Finally, it appears that late pegmatitic quartz + albite veins are particularly common in the massive greenstones. These veins do not cut across lithological contacts, and are themselves slightly folded.



Fig. 2.10

Gradation between glaucophane rich banded metabasic schists in foreground towards more massive cavernous weathering greenstones in background. Locality: Pointe des Chats (PC1). Hammer shaft is 1m long.

Sometimes the pegmatite is concentrated in large, amorphous lumps or stringers, and can abut sharply against the country rock (albitic micaschists). This phenomenon is seen particularly well at the Crique de Biléric and Inévili on the north-west coast and the Trou d'L'Enfer on the south coast.

c) Actinolite + Albite + chlorite + epidote schists ± phengite ± sphene

This is a rare basic paragenesis not previously described on Groix collected by the present author from the Crique de Biléric (specimen BIL3). This rock type differs from all the other greenschist metabasics (i.e. chlorite + albite greenstones; epidote + albite + chlorite banded schists) in having essential actinolite and albite with subsidiary amounts of chlorite and epidote. The actinolite is developed as long, thin decussate and radiating needles (up to 4 mm long, 0.05 mm wide) that make up at least 30% of the mode (see Fig. 11).

In appearance, the rock is massive, coarse and mottled green and white. The white, slightly cream weathering areas up to 5 mm wide, are albite (An < 3% confirmed by probe analysis of present author), sometimes rather inclusion poor but with the same irregular margins and untwinned character of ["]gefüllte poeciloblasts in other massive greenstones. There appears to be very little area of the rock that is not included within albite poeciloblasts, and consequently 'interstitial' chlorite is rather rare. Where chlorite is developed, it appears to form monomineralic clots of randomly orientated lamellae. Epidote occurs as isolated, highly birefringent small grains (< 0.1 mm) analysed as having ca. 21 mole % pistacite (probe analysis of present author). No clinozoisite was observed.

The present author's analyses of the actinolites compare very closely with those amphiboles analysed by Triboulet (1974) in the 'schistes vertes' of Groix and classed as 'aluminous actinolites'.



Mag x 45

Fig. 2.11

Long, thin and radiating needles of actinolite, with albite and minor clinozoisite. Specimen BIL3.

In the case of the rock BIL3, however, it is clear that bulk chemistry has had a great influence on the development of abundant actinolite, since Al_2O_3 is low (12.43 wt.%) compared with other metabasics. No whole rock or modal analyses are available for Triboulet's 'schistes vertes', so it is not possible to assess the role of bulk chemical control in the formation of aluminous actinolites in those rocks.

Seeing that the paragenesis of BIL3 is an equilibrium assemblage for the greenschist facies in basic 'sub-aluminous' rocks and that no earlier schistosity or inclusion trail relicts are preserved, it seems very likely that this rock developed during post-tectonic retrogression of pre-existing assemblages (c.f. BIL10). Sphene inclusion trails are seen, but these cut the decussate actinolites and thus must have grown very late. In outcrop, BIL3 is closely related to various foliated, fine grained chlorite + albite rocks with abundant ilmenite inclusion trails parallel to schistosity (BIL10, BIL6). It is assumed that the lack of schistosity in BIL3 points to a post-tectonic, possibly retrogressive recrystallization under greenschist facies conditions.

TABLE 4

ZONE

Phase	I	II	Retrogressive	Rock Group
Garnet				ALUMINOUS SCHISTS
Glaucophane				
Fe Chloritoid				
Mg Chloritoid		Porphyroblastic		
Chlorite				
Phengite	'sericitic'	Porphyroblastic		
Albite	'detrital'		poeciloblastic	
Rutile/ ilmenite				
Garnet	Mn-rich			BANDED EPIDOTE SCHISTS
Epidote	Granoblastic Ps=30	Porphyroblastic Ps = 20		
Glaucophane	Fe-rich	Mg-rich		
Albite				
Chlorite	Ripidolite			
Phengite		Si ⁴⁺ = 6.5 in Lawsonite pseudomorphs.		
Paragonite		" "		
Clinozoisite				
Calcite				
Barroisite				
Magnetite				
Rutile/ ilmenite				
Sphene				
Actinolite				MAGNESIAN SCHISTS
Talc				
Chlorite				
Garnet				MASSIVE BASIC SCHISTS
Omphacite				
Glaucophane				
Epidote				
Phengite		Si ⁴⁺ = 6.9		
Chlorite				
Albite			poeciloblastic	
Barrosite				
Actinolite				
Magnetite				
Rutile/ ilmenite				
Sphene				

Fig. 3.1. Facing page, pull-out diagram.

Schematic cross-section of the South Coast of the Ile de Groix, showing relationship of small scale structures to postulated megascopic folds (after Cogne et al., 1966).

- A. Overturned limb of large scale F_2 fold containing refolded F_1 axial surfaces.
- B. Small scale F_1 flat lying isoclinal folds in quartz rich layers refolded by upright F_2 folds.
- C. Highly sheared, flat lying F_1 isoclinal folds with limbs flattened parallel to S_1 .

Vertical scale is approximately equal to horizontal scale.

CHAPTER THREE

STRUCTURAL AND TEXTURAL SYNTHESIS

3.1. INTRODUCTION

The main aim of this chapter is to provide a relative time sequence of deformation events in any one area or lithology and to relate this to the various episodes of mineral growth, as deduced from textural, probe and geochronological studies.

Although it may be possible to establish a series of mineral paragenesis stable at certain points of deformational history in one outcrop, we cannot assume that the metamorphic and structural events in one body of rock are synchronous with similar events in widely separated rocks of contrasting chemical and physical properties. The difficulties of correlating fold phases with paragenetic development on Groix are made worse by the obvious changes in structural style and metamorphic grade as one proceeds from N.W. to S.E. Along the S. coast of the island, especially, there appears to be an almost complete continuum of structural styles from open upright folds near Pen Men, through recumbent folds at Gadoëric to flat-lying isoclines with boudinage at Pointe des Chats (see Fig. 1). This increase in tectonic flattening and shearing also corresponds broadly with an increase in metamorphic grade and porphyroblast grain size. Invariably, garnet-bearing metabasics are confined to an area E. of Locamria Plage on the S. coast and E. of Stanverec on the N. coast. W. of this isograd, both phengite and chloritoid in the aluminous schists occur in a sericitic habit, fine-grained and orientated planar to the dominant S_1 schistosity. By contrast, the phengites and chloritoids from the Eastern part of the island (that is, E of the isograd for garnet in metabasics) are developed as coarse grained porphyroblasts, often overgrowing and including helicitically the early

fabric elements. From this we may conclude that porphyroblastesis was in part syntectonic with the formation of S_1 in the low grade end of the island, whereas in the higher grade end crystallization was clearly post tectonic to F_1 . Conversely, glaucophane in the garnet-bearing metabasics is generally fine grained and strongly lineated compared with the glaucophanes from the non-garnet bearing metabasics which are coarser and more decussate. This latter fact might lead us to suppose that glaucophane growth was syntectonic to whichever deformation produced the lineation in the E, but was largely post-tectonic in the West.

This example of contrasting relationships of mineral texture to different deformational events serves to underline the need to define the effects of these different events on different lithologies throughout Groix. It will be necessary to set up a system of nomenclature for the various fabric elements produced by each successive deformation (viz: Table 1).

3.2.1. S_0 Original Compositional Layering

Defined in the metabasic schists by alternating epidote-rich and glaucophane-rich bands generally less than 1 cm wide. This layering was produced during an F_0 event, which brought about the mineral segregation. It is not known whether the layering is the result of original compositional banding or has come about by intense shear and transposition into parallelism of some original heterogeneity (e.g. calcareous pillow rims).

In the aluminous micaschists, S_0 is defined by quartz segregations, and possibly some rutile inclusion trails in porphyroblasts, that pre-date the formation of S_1 .

3.2.2. F_e 'deformation'

An early deformation event, possibly synchronous with F₀ but not later than F₁, that has produced a lineation followed by glaucophane needles, lawsonite pseudomorphs and pyroxene augen divergent to the b axis orientation of recognised F₁ flat lying isoclinal.

No fold closures following B_e are thought to be preserved on Groix (unless one counts the Vallon de Lavoir anticlinorium of Cogné et al. 1966), since F₁ in the east of the island has involved such a high degree of flattening and shearing. Alternatively, there is a possibility that these slightly divergent F_e lineations (generally plunging gently to 120° or 160°) were produced during F₁ by rotation of the isoclinal axes, which may vary widely about 160°, due to non-cylindricity of folding.

3.2.3. S₁ schistosity

S₁ schistosity in the metabasics is often intense, planar and parallel to the axial surfaces of F₁ isoclinal, and is the plane within which the mineral lineation is contained (see Chapter 2, Fig. 8; of LM12 locality). In the micaschists, S₁ is a schistosity composed of a planar orientated mica fabric. This principal planar schistosity is often parallel to lithological contacts and to the axial planes of flattened quartz rodding.

3.2.4. F₁ deformation

F₁ is the major deformation episode on Groix, and has produced quartz rodding in the micaschists and isoclinal folding of epidote layers in the metabasics. Quartz rods were formed by the shearing out of flat lying isoclinal in S₁, and generally indicate a westerly

'tectonic transport' direction parallel to S_1 , with elongation directions plunging gently to $160-180^\circ$.

F_1 isoclines in the metabasics are of two types:

- a) metre scale overfolds with rounded hinges and small scale drag folds as markers
- b) cm scale flat lying isoclines with sharp apices and sheared out limbs.

It is type a) (see PHOTO Fig. 2) which demonstrates how relatively competent 2-6 m metabasic horizons can flow parallel to S_1 during F_1 deformation, and eventually form boudins by necking of the fold limbs (see Fig. 3).

In the east of the island, where S_1 is nearly horizontal, boudinage is very common amongst the metabasic schists, and the axes of elongation are parallel to both B_1 fold axes and mineral lineations.

Further west on Groix, it appears F_1 deformation was not so intense, since mineral lineation (parallel to 'b' axes) and boudinage are less marked. Also, the trend of F_1 folding swings round, as one approaches Pen Men in the N.W., to assume a NW-SE orientation. However, F_1 folds are orientated NE-SW on the high grade part of the north coast west of Port Tudy, and are picked out by epidote bands which show the isoclines to be due to the first deformation of S_0 , with their axial planes parallel to S_1 .

3.2.5. F_2 deformation

F_2 deformation on Groix produces open, upright flexures in S_1 on the mesoscopic scale in the micaschists, with fold axes plunging gently to $140-160^\circ$. On the small scale, a crinkle lineation is formed on micaceous surfaces where the S_2 vertical strain slip cleavage intersects,



Fig. 3.2

Type a) F_1 isocline in banded epidote schists. Locality: Pointe de la Croix. Compass is 10cm long.



Fig. 3.3

Metabasic boudins encased in chloritoid micaschists. Looking north; glaucophane lineation in boudins is N-S. Locality: E. side Por'h Morvil. Open compass is 20cm long.

and refolds, the S_1 schistosity.

Sometimes, where F_1 and F_2 fold axes are sufficiently divergent, interference patterns are created which have been described by Ruhland and Jeannette (1966) under the general term of 'fuseaux'. These 'fuseaux' are particularly abundant in graphitic micaschists and phyllites on the south coast of Groix between Por'h Roëd and Pointe de L'Enfer. Here, F_2 deformation can locally be very intense, as witness the NE dipping shear planes at Ker Bethanie which have controlled the refolding of the 'fuseaux' (see PHOTO Fig. 4).

On the large scale, the most important effect of F_2 is the refolding of S_1 where F_1 and F_2 are co-axial. This is particularly well observed at the Port St. Nicholas in the form of an upward converging fan of S_1 schistosity within a $\lambda = 500$ m anticlinorium of greenstones and phyllites.

F_2 deformation may also be responsible for the changes in S_1 attitude from Locmaria to Pen Men, though this change from flat-lying to upright S_1 may in part be due to a varying style of F_1 deformation.

As regards the metabasics, F_2 deformation has produced a vertical strain slip cleavage S_2 , composed of $\lambda = 2-5$ cm open folds plunging gently to $130 - 160^\circ$. In the areas of the island where F_1 and F_2 are almost co-axial, 'b' glaucophane lineation is not bent, but the axial surfaces of the isoclines (S_1) are buckled (PHOTO Fig. 5). Elsewhere (as at Plage du Trec'h, Kermarec and Gadoëric) glaucophane lineation is clearly refolded about upright F_2 crenulations which plunge gently to 160° .

F_2 crenulations are distinguishable from F_1 isoclines only when there is either a difference in 'b' axis orientation or axial plane dip. Since the orientation of F_1 axial planes and fold axes is demonstrably variable across Groix, it follows that there can be areas where, on the outcrop scale, F_1 and F_2 are indistinguishable in their effects. This has happened at Gadoëric, where an early F_e fine grained glaucophane



Fig. 3.4

Interference 'fuseaux' in graphitic micaschists. Horns of 'fuseaux' point NW-SE, whilst crinkle lineation in centre of body is N-S. Locality: Baie des Curés. Scale is 5cm.



Fig. 3.5

Epidote rich bands pick out a refolded F_1 isocline. Rectangular pseudomorphs after lawsonite concentrated in band near centre of photograph. Locality: Stanverec. Compass is 10cm long.

lineation, divergent to local isoclinal axes, has been refolded by a later upright fold phase trending $160 - 340^{\circ}$, which could be either F_1 or F_2 depending on the relationship of the S plane to larger scale structures.

This particular case highlights the difficulty of correlating fold phases where

- a) the structural style is variable from area to area within one lithology.
- b) there is a lack of inland exposure and vertical cliff exposure is restricted to 30m in accessible areas.

Because of the above factors, it has proved very difficult to delineate the geometry and attitude of large scale structures on Groix. So it has to be assumed that the F_1 folds seen on the outcrop scale are of the same generation as the principal planar schistosity S_1 .

Fig. 1 shows a 3 km long schematic cross section of the postulated large scale structures (Cogné et al. 1966) in relation to S_0 or S_1 observed on the outcrop scale.

3.2.6. F_3 deformation

F_3 is a deformation which has produced an important set of E-W trending kink bands and $\lambda = 5$ cm chevron folds in micaschists throughout Groix and mesoscopic, open cross-folds trending 80° - 260° in metabasic horizons for the Port Méhite - Pointe de Spernec sector. The kink bands and axial surfaces of F_3 chevrons define an S_3 strain slip cleavage (see PHOTO Fig. 6) which is variably developed in the micaschists, but generally concentrated in several 'belts' adjacent to faults. Both fault planes and S_3 dip steeply to the north, and are associated with a conjugate joint set (40° , 100°) which suggests that maximum principal stress was directed horizontally N-S during F_3 and that deformation was brittle in character. In addition, tension gashes orientated E-W probably opened



Fig. 3.6

F_3 strain slip cleavage in micaschists. Locality: Pointe des Chats, Pen Ganol.

up during F_3 and were subsequently filled with quartz-albite or chlorite-albite pegmatites (vein and cavity-filling rocks of Makanjuola, 1967).

F_3 deformation has folded 2-6 m metabasic horizons into open flexures ($\lambda = 2-5$ m) of variable amplitude but all trending 80° - 260° , in an area where the effects of F_2 folding are very weak and a marked band in megascopic S_1 strike from N-S to NE-SW is apparent. This bend in schistosity strike is parallel to the bend inferred for the 'garnet in metabasics' isograd (see Fig. 1, Chapter 2), and may be due to a megascopic external rotation of unknown generation about an ENE-WSW axis plunging steeply to the WSW.

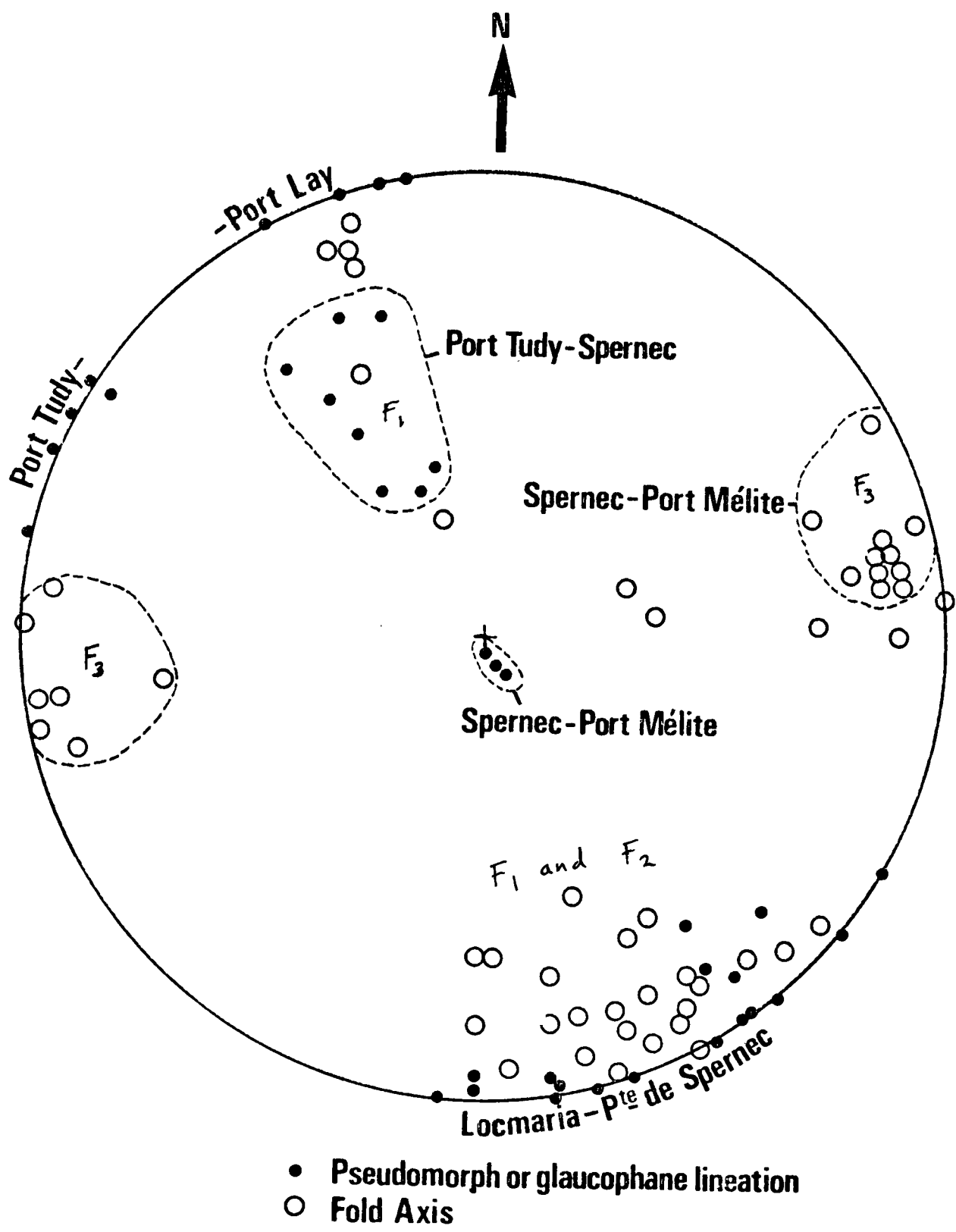
In this sector, also, F_1 glaucophane and lawsonite pseudomorph lineations have been refolded away from the primitive about an axis typical of F_3 (see Fig. 7). If we identify the mesoscopic F_3 folds with the constructed β pole of the rotated S_1 , then F_3 deformation must be post prograde metamorphism and possibly syn retrogression, since the tension gashes associated with F_3 faulting and folding are filled with late-stage pegmatitic albite, quartz and chlorite, and cut all pre-existing S planes.

Another effect of F_3 deformation is the production of a fine crinkle lineation trending E-W that cuts across F_3 crenulations in graphitic 'fuseaux'.

3.2.7. F_4 deformation

F_4 is a deformation that is only recognised by its effect on albite orientation fabric in retrogressed greenstones. Between the Sémaphore house and the Crique de Biléric is a small ledge where Makanjuola (1967) collected his magnesian schists. It is at this locality (PHOTO Fig. 7 in Chapter 2) that large albite euhedra in a chlorite + albite greenstone band show a distinct orientation parallel to the axial

Fig. 3.7 Stereographic projection (into S. hemisphere).



plane of NE-SW trending folds. This axial planar S_4 is clearly post-retrogression, which elsewhere has been inferred to have taken place after F_2 and during F_3 . Although weak albite orientation is observed elsewhere on Groix, especially in aluminous albitic mica-schists, it is never clearly parallel to the axial surface of observed mesoscopic folds.

It is possible that other NE-SW trending cross-folds and kink bands belong to the F_4 phase, but the critical evidence of cross-cutting or interference relationships is lacking. F_4 deformation at Biléric has produced rather irregular contortions in S_0 , since all other planar fabrics have been obliterated here by retrogression. However, the adjacent magnesian schists at Biléric seem unaffected by F_4 deformation since they show decussate, random orientation of grains. This suggests that the magnesian schists recrystallised post-tectonic to F_4 and thus very late in the deformation history of Groix.

3.3. Geographical distribution of tectonic style

The two main fold generations on Groix, F_1 and F_2 , not only differ in style from one another but also are very variable in B axis orientation and axial plane attitude from area to area.

F_1 is most strongly developed in the eastern part of the island (see Fig. 1) from the Vallon de Storang on the south coast through to Pointe des Chats and round to as far as Inéveli on the NW coast. Throughout this area F_1 deformation has given rise to small scale isoclinal folds in epidote banded schists and quartz rodding in micaschists (see Table 1) but the axial planes S_1 vary in attitude from flat-lying (near Pointe des Chats) to dipping steeply to the NE (near Port Lay). Going from SE to NW, the F_1 fold axes swing round with shallower plunges to 180° - 140° . This variation seems systematic and continuous from SE to

TABLE 1

Graphitic schists and phyllites

Micaschists

Metabasics

F ₀	Original ? sedimentary layering	Segregation into epidote and glaucophane rich layers. Some early (F _e) mineral lineation.	
F ₁	Quartz rodding produced from sheared out isoclinal S ₁ composed of planar mica fabric.	Flat lying isoclinal and boudins with b plunging gently to 180°-160° mineral lineation and boudin elongation generally parallel to b. Folds face W.	F ₁ isoclinal only preserved in cores of interference 'fuseaux', with axes N-S and flat lying axial planes
F ₂	S ₁ folded into open, upright flexures, plunging gently to 140-160°. Small scale crinkled lineation // vertical strain slip cleavage S ₂	F ₁ isoclinal refolded by vertical slip λ = 2-5 cm open folds, producing strain slip cleavage, S ₂ . On large scale F ₂ refolds F ₁ axial planes and produces open folds in greenstone horizons.	S ₂ shear planes dipping 50° to NE with associated drag folds plunging to 160° and refolding horns of graphitic fuseaux (Ker Béthanie)
F ₃	Kink bands and λ = 5 cm chevron folds produced, trending E-W, axial planes dipping steeply to N.	Mesoscopic open, cross folds trending E-W in the Port Melite-Pointe de Spenec sector.	Fine crinkled lineation in graphitic layers, trending E-W.
F ₄		NE-SW trending cross folds at Bileric that orientate albite // S ₄	

NW and a decrease in metamorphic grade occurs towards the "Faltenspiegel" (fold mirror plane, as defined by Felix, 1969, between sub-areas dominated by F_1 or F_2 folds).

It is interesting to note that the postulated trace of the "Faltenspiegel" is exactly parallel to, and offset to the W or SW from, the trace of the 'garnet-in-metabasics' isograd observed by the present author (see Fig. 1, Chapter 2). This would tend to suggest that a prograde metamorphic sequence was established very early in the deformational history and that the resulting isogradic surface was bent on a regional scale before F_2 folding.

By the time F_2 deformation was in operation a large scale bend already existed in the F_1 structures, and this to some extent controlled the style, direction and intensity of F_2 folds. (To the west of La Baie des Curés on the south coast and to the west of Inéveli on the north coast is the area where F_2 is the most important phase of deformation. This may in part be due to the fact that metabasic horizons are less common and thus the chances of preservation of F_1 folds are smaller).

In this area west and south-west of the "Faltenspiegel" the folding is open in style (30° interlimb angle) on the mesoscopic scale, and the axial planes all dip steeply to the SW. On the small scale, F_2 expresses itself as an upright strain slip cleavage that refolds epidote isoclinal and quartz rods. The B_2 axes vary from 130° - 160° , all with shallow plunges, as one goes from W to E, which rather parallels the bend in S_1 schistosity strike though not to such a marked extent. From this last fact one may conclude that F_2 deformation about broadly NW-SE axes was not responsible for all the changes in F_1 fold and schistosity orientation that are observed. However, an inspection of stereograms presented in Cogné et al. (1966) will show that F_1 axial planes have suffered rotation about mesoscopic B_2 axes, especially in the region immediately to the east of La Pointe de St. Nicholas. But on the larger scale,

the changes in S_1 attitude cannot be accounted for by having one simple post F_1 fold generation. Rather, the apparent rotations observed could be due to variations in the F_1 stress fold reflecting lithological inhomogeneties on the regional scale. If this were so, then there would be no reason to suppose that isogradic surfaces generated during metamorphism should necessarily be parallel to early formed schistosity (S_1). This difficulty can be overcome if we postulate that the variations in F_1 fold axis orientation and S_1 attitude are due to an external rotation during or directly after F_1 , but before F_2 . Then during F_2 , F_1 axial planes were folded and B_1 axes rotated only in those areas where F_2 deformation was of sufficient intensity, i.e. west of the 'Faltenspiegel'.

3.4. Cogné's interpretation of the structural-metamorphic history of Groix.

According to Cogné et al. (1966), Groix affords a unique opportunity to observe the effects of Hercynian folding and greenschist/prasinite facies metamorphism superimposed on an already highly deformed and metamorphosed basement of presumed Cadomian age (690-570 M.yr, C.J. Adams, 1967). The contrasting fold styles and metamorphic assemblages developed in these separate orogenic phases are summarised in Table 2.

Cogné et al. (1966) consider that the earliest folds preserved on Groix, characterised by flat lying isoclinal with round hinges and B axes now plunging gently to either 180° or 120° , were flexured, flattened and stretched out during the 'Bretonic' phase of deformation.

During this latter deformation, which we shall for the sake of argument term F_B , an almost complete metamorphic recrystallization took place in a facies intermediate between greenschist and albite epidote-amphibolite ('prasinite') facies. In the boundary zone between these two facies, a glaucophane schist facies was developed where the NW-SE Hercynian folds interfered with the earlier E-W Cadomian structures

TABLE 2. COGNE'S (1966) INTERPRETATION OF THE STRUCTURO-METAMORPHIC HISTORY

Metamorphic Assemblage			Tectonic style
Basic		Pelitic	
Omphacite + Garnet + Sphene ? Kyanite Hornblende + Garnet + ? Plag		Biotite + Muscovite + Chlorite ± Garnet	Flat lying isoclinal ¹ , overturned to S or SW, with rounded hinges. B axes plunge gently to 120.
Glaucophane		Chloritoid	Flattening of F ₁ isoclinal; Boudinage. Intense parallel planar schistosity produced (S _B)
Blue green amphibole Albite Chlorite Magnetite			NW-SE upright, open folds E-W kink bands

F_A
? Cadomian

Hercynian { F_B Bretonic phase
F_C Sudetic phase

to produce N-S isocline axes and mineral lineations. Thus a syn M_B recrystallisation during F_B flattening of early isoclinal led to an intense planar lamination parallel to the axial planes. Boudinage during this F_B phase was produced by N-S stretching, as opposed to the F_A phase which gave rise initially to E-W stretching of the boudins (see Fig. 8).

Thus F_B in the SE is expressed by mesoscopic overfolds whose cores are not actually observed in outcrop. Instead, the intense parallel planar schistosity (S_B) associated with these folds is very apparent, and boudin elongation axes are parallel with the dominant N-S syn-tectonic glaucophane lineation.

As one proceeds west, the style of F_2 changes to more upright NE-SW isoclinal whose axial planes are clearly refolded by F_3 , a fold phase characterized by open style and S.W. dipping axial planes. F_C deformation is thought to correspond with the sudetic phase of the Hercynian orogeny, and is accompanied by a retrogression of the prasinites and glaucophane schists which increases in intensity towards the west.

3.5. Objections to Cogné's model

The present author could cite much petrographic, geochronological and microprobe evidence that is not in accord with Cogné's model for the metamorphic evolution of Groix, but here proposes to compare the sequence of deformational events as presented in Tables 1 and 2 with those in Table 3, and point to the lack of evidence for a pre-glaucophane schist ? Cadomian metamorphism. The evidence regarding phases of mineral growth is discussed in more detail in the subsection titled 'Mineral growth episodes in relation to deformational history'.

No evidence has been found during extensive mineralogical studies of the Ile de Groix (Makanjuola & Howie 1972, Triboulet, 1974, this study) for any relict primary biotite from the pelitic schists or horn-

blendes from the basic schists. The biotite that has been tentatively recorded on the Ile de Groix (Triboulet, 1971) falls into two parageneses, neither of which is primary in the sense of being pre F_2 .

- a) Brown biotite possibly in equilibrium with garnet in pelites of Triboulet's (1974) highest temperature zone.

Also associated in interbedded quartzose feldspathic schists is stilpnomelane + brown biotite.

- b) Green biotite is widespread as an accessory in a wide range of different lithologies, but is always secondary after prograde metamorphic minerals and probably grew during retrogression.

In contrast with the biotite + muscovite chlorite schists or quartzose schists of Le Pouldu Series on the mainland, which are thought by Cogné to preserve pre-Hercynian metamorphic assemblages, the quartzo-sericitic schists of Groix contain no primary biotite. Similarly, there is no trace of residual hornblende left over from Cogné's postulated primary amphibolite facies metamorphism. The only green amphiboles analysed are aluminous actinolites from highly retrogressed magnesian schists or chlorite-albite greenstones (Makanjuola & Howie, 1972, Triboulet, 1974, this study). Furthermore, the zoning profiles of glaucophane never show grain cores approaching hornblende in composition (this study).

Again, this is in contrast with Le Pouldu Series, where a hornblende ± epidote + chlorite + albite assemblage is typical for the basic rocks, and no trace of glaucophane, garnet or omphacite has been observed.

Cogné also quotes several examples of garnet retrogression, and assigns the breakdown of garnet to the 'Bretonic' phase. This is directly in contradiction to the idea of glaucophane being in equilibrium with garnet during a late stage of the prograde metamorphism; an idea

put forward and defended by the present author in the succeeding sections.

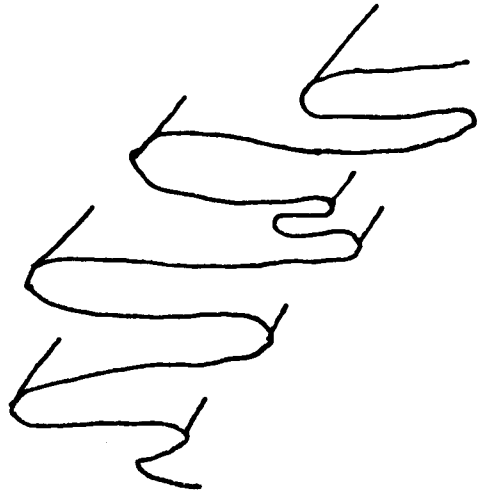
As far as structural history is concerned, the present author would shrink from attaching orogenic phase labels, such as 'Bretonic', 'Sudetic' to various deformational phenomena in rocks lacking fossils or recognizable unconformities. Instead, a relative time sequence of deformational 'events' in rocks of differing grade and lithology is defined, (as in Tables 1 and 3), correlated across the area concerned (as in the section on geographical variation in tectonic style) and used as a framework for textural interpretation. The end-product of this methodology is the recognition of a series of mineral growth episodes, whose significance and duration is explored in the chapters on conditions of metamorphism and geochronology respectively.

The present author sees no reason at this stage of the argument to invoke separate deformation phases to account for

- a) flat lying isoclines with rounded hinges and E-W stretching of boudins
- b) flat lying small scale isoclines with angular hinges, flattened limbs and N-S elongation of associated boudins and mineral lineation.

Since Cogné assigns a) to the earliest, pre-glaucophanization deformation and metamorphism (Cadomian), he is forced to the conclusion that the intense glaucophane/epidote banding (defined as S_0 by the present author) was produced during M_B metamorphism but after F_A deformation. This glaucophane must have grown syntectonic to his F_B (Bretonic phase), which flattened and rotated F_A isoclines parallel to S_B . In the stress field so generated, glaucophane would have grown with their long axes along the minimum principal stress direction (σ_3) of these folds.

However, both type a) and type b) isoclines could have formed in the same triaxial stress field during the same deformation (F_i as defined by



Type a) isoclinal



Type b) isoclinal

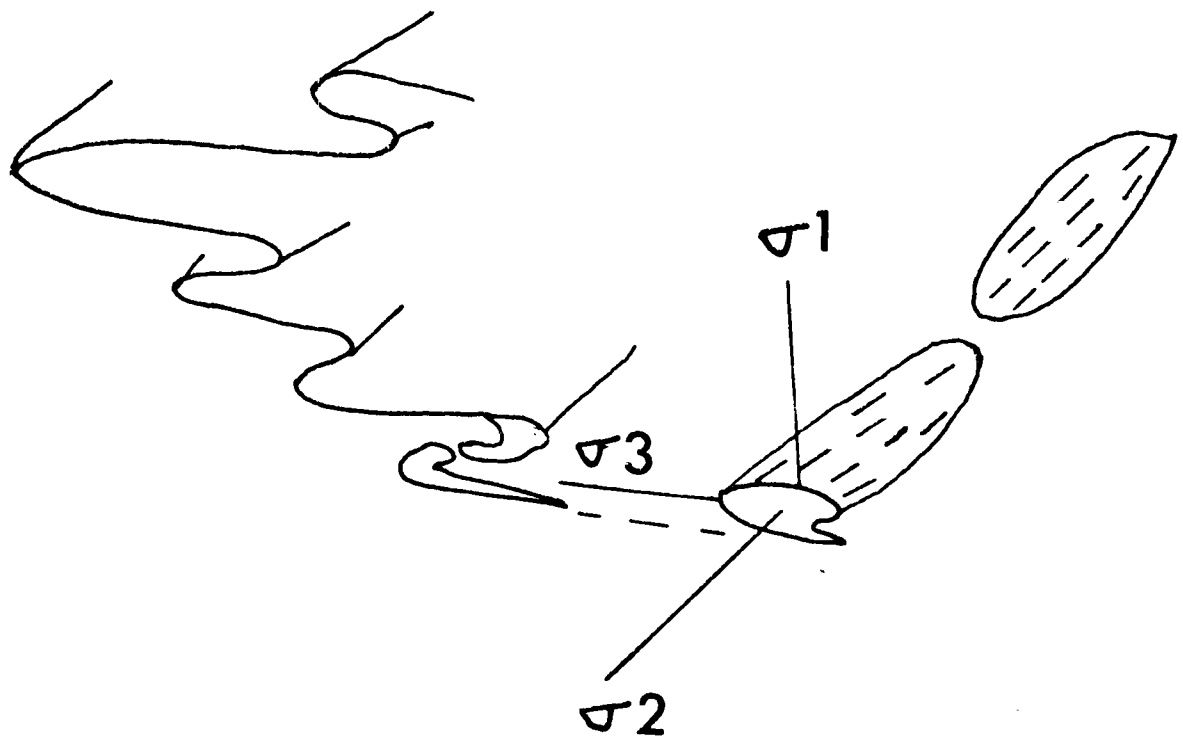


Fig. 3.8 Model for production of boudins in triaxial F_1 stress field. Boudinage occurs by stretching of more brittle (metabasic) horizon in a ductile medium (micaschist). Glaucophane, lawsonite and garnet augen are aligned parallel to σ_2 after separation of the boudins.

the present author) if σ_3 corresponded with the E-W 'boudin necking' direction and σ_2 with the N-S mineral lineation and 'boudin elongation direction' (see explanatory Fig. 8).

In this way, one need have no recourse to early phases of deformation that rely on divergent mineral lineations or presumed (not observed) b axes for their definition. Any variations of F_1 fold orientation across the area can be accounted for quite simply by either external rotations during F_1 or refolding locally by F_2 .

3.6. Mineral growth episodes in relation to deformational history

This section attempts to combine the observed textural relationships between co-existing planes with the account of tectonic fabric development given above. The aim is to achieve a relative time scale for mineral growth on Groix, against which various hypotheses of paragenetic development can be tested. Briefly, there are two contrasting ways of explaining the sequence of mineral growth as inferred from thin section studies. The first, propounded by Mäkinen in 1967, seeks to explain the supposed prevalence of disequilibrium assemblages in terms of successive, and partial retrogressions from an early eclogite facies assemblage.

Mäkinen (1967) concluded that "in Groix, metamorphic changes during each of the episodes took place at relatively lower temperature than during the preceding one and this has given rise to arrested reactions by which older parageneses are carried over and preserved among the newly produced mineral assemblages". The episodes that Mäkinen refers to here are those of mineral growth (viz. pre-glaucophanization, glaucophanization and post-glaucophanization) and bear no relation to deformation.

In contrast, Felix (1969) and Triboulet (1974) recognise a primary prograde metamorphic sequence that was produced synchronously followed

later by a 'patchy' retrogression into the greenschist facies.

Again, there is no attempt to relate mineral growth with deformation episodes.

Table 3 is here presented as a summary of the present author's views on the sequence of mineral growth, as deduced from tectonic fabric and thin section observations. This table is compiled from the petrographic observations in Chapter II of Makanjuola's 1967 thesis, from Felix's (1969) petrographic study of the basic rocks and from Felix's (1972) structuro-mineralogical study of lawsonite pseudomorphs. The present author has added his own thin section and microprobe observations to the above data and has interpreted the mineral growth chronology so inferred in terms of the deformational history laid out earlier in this chapter.

The textural observations of the present author agree fairly well with those of Makanjuola, though the interpretation in terms of paragenetic development is quite different. Basically, the present author considers that there is no evidence for the glaucophanization episode postulated by Makanjuola to explain the addition of glaucophane to pre-existing eclogitic assemblages. Furthermore, the garnet + omphacite + sphene assemblages are thought to have developed post-lawsonite breakdown in the basic schists.

Generally, the present author is in agreement with Felix (1972) and Triboulet (1974) as far as the recognition of a primary prograde metamorphic sequence is concerned, but has broadened the argument to include microprobe and geochronological data.

In the succeeding sub-sections, the main points of textural interpretation, upon which the author's conclusions are based, will be discussed.

3.6.1. Evidence from rutile/sphene relationships for a prograde (post F_1) development of garnet + omphacite assemblage in dehydrated zone I metabasics

Makanjuola himself admits in his 1967 thesis (p.98) that sphene/rutile relationships in the pyroxene bearing basic schists are the reverse of what he would expect if his hypothesis of glaucophanization of sphene bearing eclogites were correct. Always sphene is seen to rim rutile, and garnets contain rutile inclusions when not associated with omphacite.

If rutile growth is assigned to an early episode (pre or syn F_1) because of the widespread occurrence of rutile inclusions trails parallel to S_0 or S_1 in almost every orientated porphyroblast phase, then sphene must have grown subsequent to F_1 deformation.

In the pyroxene bearing glaucophane schists (i.e. present author's PC10, LM12), sphene is the only mineral included in the omphacites. From the fact that poeciloblastic garnet includes sphene, epidote and omphacite, one can assign a very late stage of the prograde metamorphism for the growth of garnet (see sub-section 11.4). That being so, then the equilibrium temperatures and pressures derived from the garnet-omphacite-quartz-albite assemblage in Chapter 4 of the present study (400°C, 8 kb) must also refer to this late stage of the prograde metamorphism, when early-formed rutile and lawsonite had broken down.

If progressive metamorphism was a history of increasing temperature and dehydration, one would expect to form omphacite at the expense of glaucophane viz:



Thus sphene, if included in the omphacite must have formed from rutile before the above reaction took place but after lawsonite breakdown, since small euhedral garnets in lawsonite pseudomorphs have rutile inclusions

whereas the pseudomorphing material contains sphene (see sub-section 11.7).

Unlike New Caledonia, where sphene is stable in the lawsonite (low grade) zone and rutile in the epidote (high grade) zone, Groix has rutiles often rimmed by sphene and sphene is the only stable Ti-bearing phase in the high grade (Zone II) pyroxene bearing glaucophane schists.

This different facies series vis-a-vis sphene/rutile relationships could be explained by a smaller (ca 12°C/km) P/T gradient for the New Caledonian blueschist metamorphism than for Groix (ca 16°C/km) (P/T gradient estimates from Chapter 4 on conditions of metamorphism).

In some rocks on Groix, especially those from the non-garnet bearing zone I, sphene must have grown post F_2 and contemporaneously with retrogression, because sphene is the only Ti-bearing phase in totally recrystallized chlorite + albite greenstones and is never seen as inclusions in relict F_1 or F_2 porphyroblasts.

3.6.2. Evidence for primary syntectonic growth of glaucophane

Glaucophane can occur as fine (0.25 - 1 mm) lineated blades or as large (1.5 cm) stubby decussate laths. The former habit is commonly developed in the metabasic schists of zone II, where it appears the glaucophane grew in a stress field that produced the F_1 isoclinal overfolds and boudins since the z crystallographic axes are preferentially orientated in the foliation plane (S_1) parallel to the b axes of observed F_1 isoclines. Occasionally, glaucophane lineation is divergent to the b-direction of F_1 isoclines, and this has been interpreted by the present author as being the result of syntectonic crystallization to F_e , an early isoclinal fold phase, the fold cores of which no longer exist.

Whatever the relation of glaucophane lineation to observed fold axes, the grains are never broken or strained. Even when F_2 open folds

have quite clearly refolded the F_e or F_1 lineation, glaucophane grains show no sign of strain or chemical inhomogeneity. This can be interpreted as post- F_1 mimetic crystallization due to post-deformational heating. However, if temperatures during metamorphism are high enough and strain rates are fast enough, it is very likely that most crystalline solids should show evidence of recovery and recrystallization (Vernon, 1976). There is no reason, in the case of the Groix glaucophanes, why annealing and the removal of incremental strain could not occur continuously during deformation, thus leading to the growth of polygonized sub-grains. In this way glaucophane, though formed early in the deformational history, would continue to recrystallize and escape the effects of cumulative strain. The idea is far more consistent with an early lawsonite + glaucophane paragenesis on Groix (see 3.6.5) and obviates the necessity of appealing to a glaucophanization process (Makanjuola and Howie, 1972).

The latter mentioned habit of glaucophane growth (see PHOTO Fig. 13) is typical of non-garnet bearing metabasics and aluminous schists from zone I, where F_2 deformation is important (e.g. Stang er Marc'h, Ker Bethanie). Here, the glaucophanes are only crudely orientated within S_1 , and have especially Fe-rich compositions (e.g. PM6, KB1, PNI). It is thought by the present author that ferro-glaucophane growth continued unchecked during F_2 deformation in zone I, whilst in higher grade rocks, where F_2 was not so intense, garnet grew instead, in equilibrium with more Mg-rich glaucophanes.

3.6.3. The relationship of glaucophane to garnet and omphacite

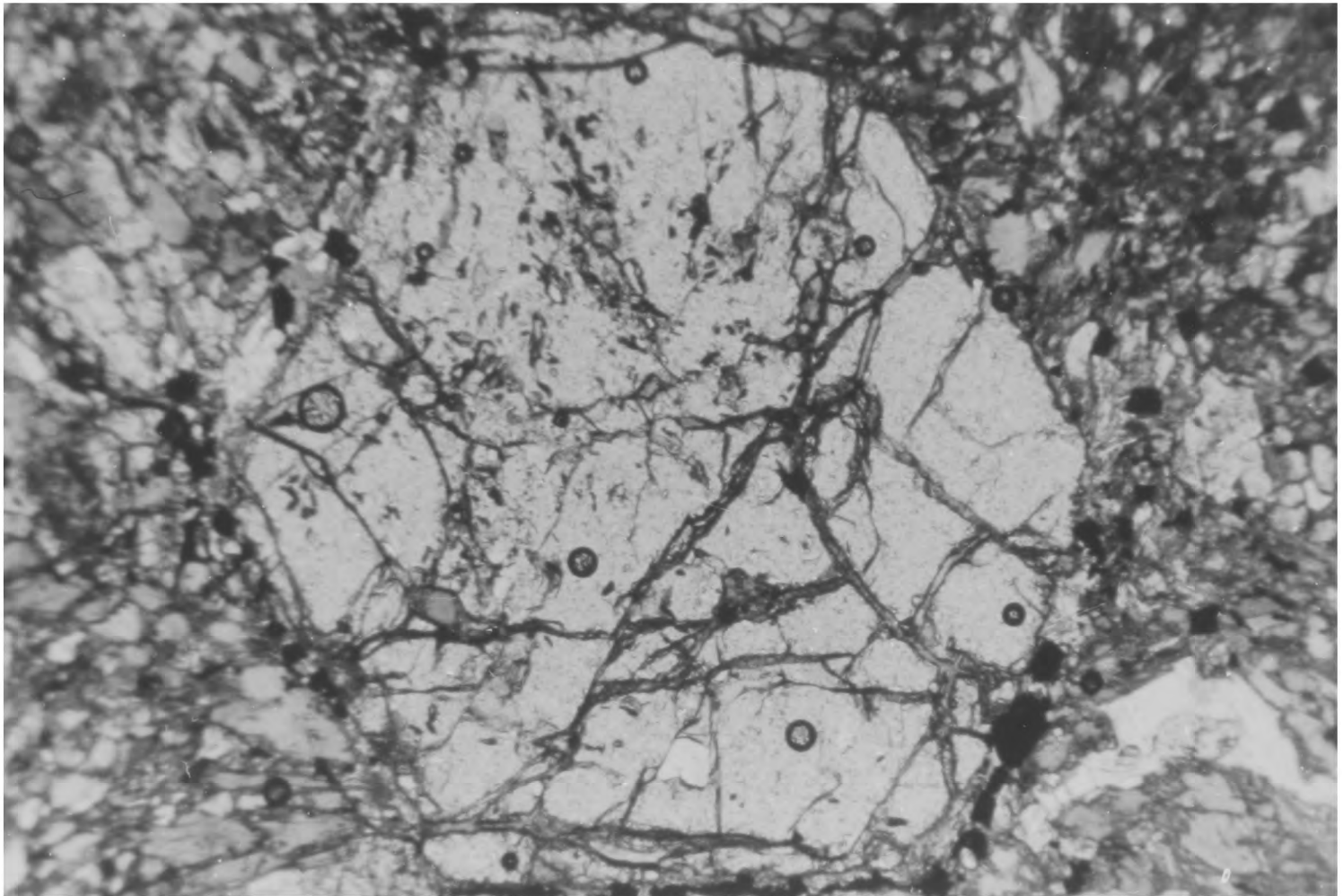
Makanjuola and Howie (1972) have reconstructed the mineral parageneses of basic schists on Groix in such a way that garnet never appears in equilibrium with glaucophane. They have interpreted the omphacite + garnet + sphene assemblage as being primary, in contradiction to much of

the textural evidence presented in the present study, and have assigned glaucophane growth to a "late paragenetic re-adjustment under waning temperature conditions".

In thin sections of pyroxene-bearing glaucophane schists the present author has observed sharp contacts between unzoned glaucophanes and garnet rims (PHOTO Fig. 11 of PC10), which shows that both minerals remained in equilibrium after omphacite and sphene began to grow, since omphacite and sphene are commonly included in this type of garnet.

Furthermore, glaucophane is often included in garnets from omphacite-free, epidote banded glaucophane schists (e.g. PHOTO Fig. 9 of TUD5) indicating that glaucophane pre-dated garnet growth in some cases. This latter observation is in accord with the present author's belief that some glaucophane growth on Groix was early in relation to garnet or omphacite growth, and probably synchronous with lawsonite and rutile development. The microprobe analyses (see Appendix) of such included glaucophanes show an enrichment in Fe compared with glaucophanes in the groundmass, and this can be interpreted as a relict from the lower grade, non-garnet bearing prograde assemblage.

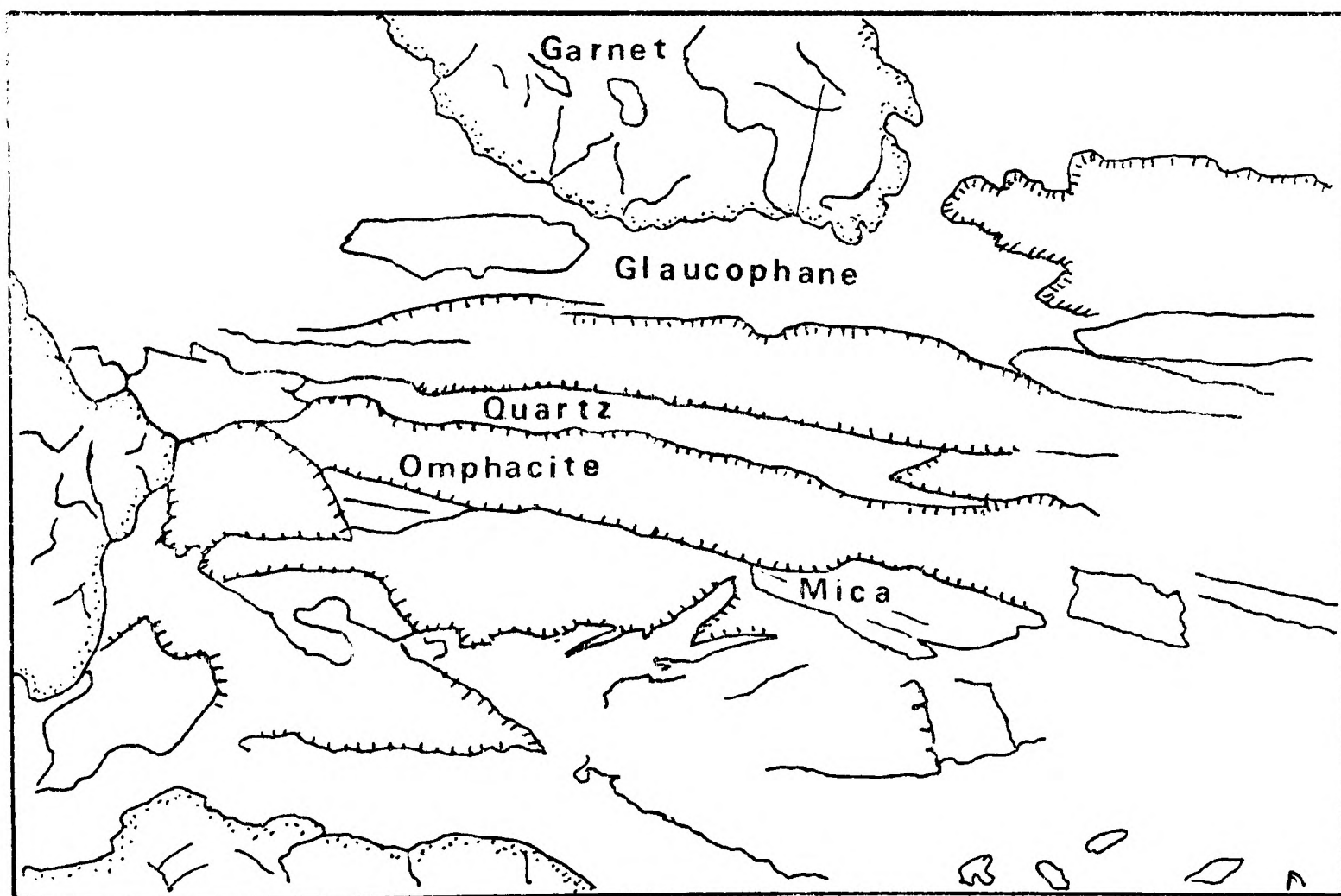
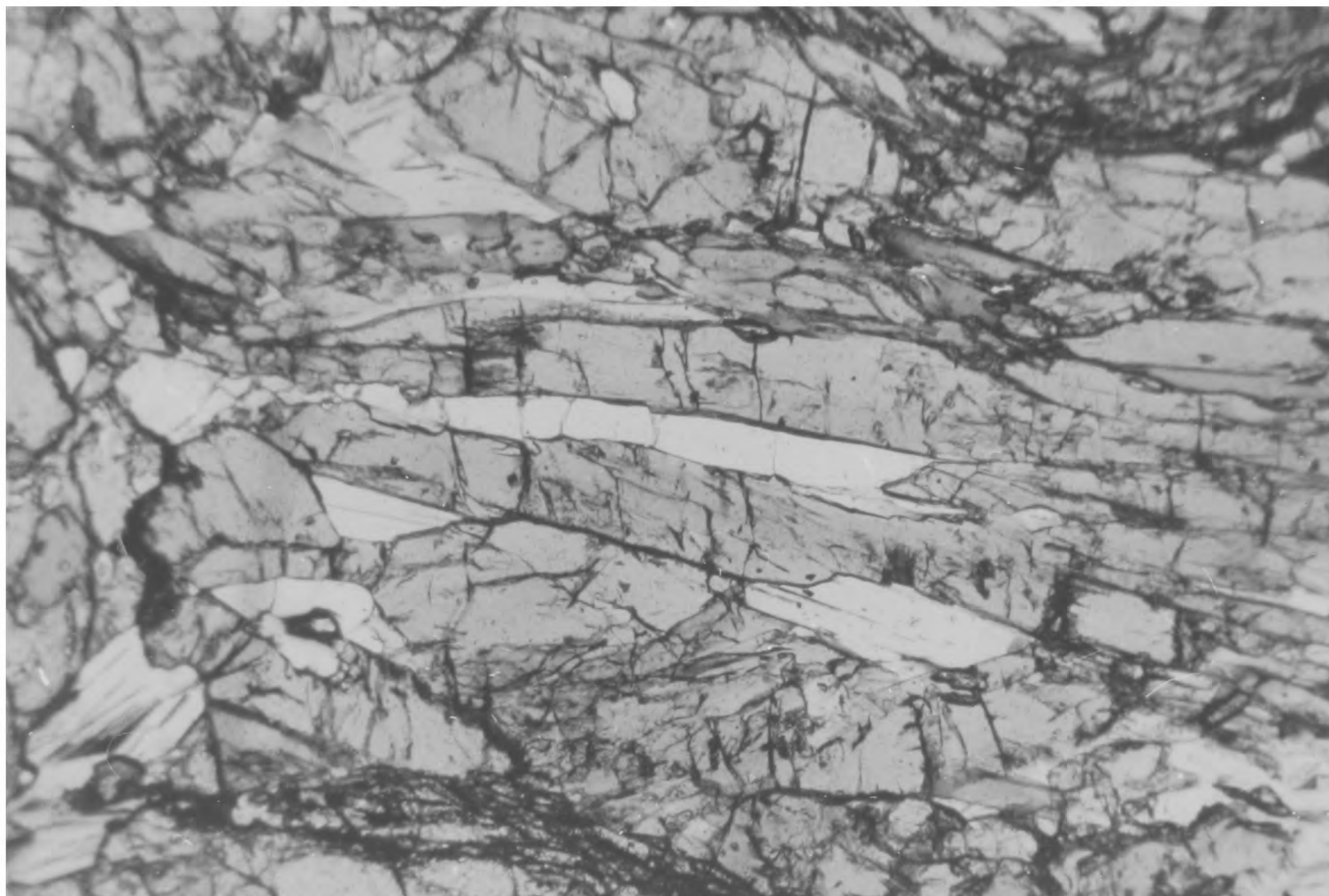
Although glaucophane in the pyroxene bearing schists is clearly in equilibrium with garnet rims, and the garnets have every appearance of being in equilibrium with the adjacent omphacites, the contacts between glaucophane and omphacite are always filled with a screen of barroisite (see Appendix III for analyses of PC10 barroisites, and PHOTO Fig. 3.11 for juxtaposition of spots). Thus it is not possible to say whether glaucophane pre- or post-dates the growth of omphacites in these rocks if one just considers this observation in isolation. However, if one combines the evidence from sphene/rutile relationships (subsection 3.6.1) with the evidence for mimetic growth of the pyroxene augen one must conclude that omphacite developed at the expense of



Mag. x 45

Fig. 3.9

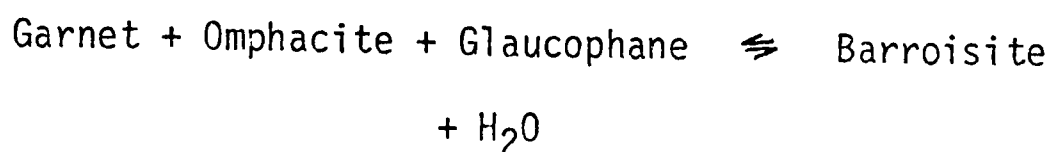
2.3mm euhedral porphyroblast of garnet with inclusion of ferro-glaucophane in bottom left hand corner. Cracks in garnet are filled with chlorite. Other inciusions are rutile and epidote.



mag. X 45.

Fig. 3.II Pyroxene augen in glaucophanic eclogite PCI0.

glaucophane, and possibly epidote, in areas of the rock that had 'dried out' during prograde metamorphism. That being so, then barroisite must have formed later when water became available again during retrogression:-



3.6.4. Evidence for post-tectonic garnet growth

Inherent in Mäkinen's scheme of successive mineral replacement is an assumption that the 'eclogitic' garnet + omphacite pre-dated glaucophane growth and hence should be pre-tectonic to the deformation that orientated the glaucophane. If this was so, garnet should show signs of rolling or cataclasis.

However, the present author believes there is no conclusive evidence for an important phase of pre-tectonic garnet growth in the metabasic rocks of Groix. Inclusion swirls of epidote, rutile or sphene in some garnets can be interpreted as F_1 folds helicitically overgrown by post F_1 garnets rather than being produced by rotation of the garnet during growth (Spry, 1969). Similarly, the occurrence of spongy, inclusion-rich garnets with a fish-net or web texture (spec. PC5) is highly suggestive of post-tectonic garnet crystallization along the grain boundaries of a polygonal quartz mosaic. In this case, the quartz recrystallized into an unstrained mosaic post-tectonically and garnet grew over the mosaic.

In the pyroxene bearing metabasic schists, 5 mm garnet euhedra quite often form the nuclei of augen composed of unstrained omphacite, quartz and phengite (e.g. PC10). These augen are elongated in a direction parallel to glaucophane lineation, but divergent to F_1 isoclinal b-axes (c.f. LM12 locality) and F_2 crenulations in micaceous interlayers. Thus it appears that the augen minerals have crystallized mimetically to

an F_e lineation, since they are unstrained and show 'equilibrium' contacts with post F_1 garnets. It is considered unlikely that these omphacitic augen represent relict eclogitic assemblage preserved in the 'pressure shadow' of garnets, since firstly the garnets are demonstrably post-tectonic and secondly there is no trace of F_1 deformation affecting these augen. A more likely explanation is that locally glaucophane and epidote, which followed an F_e lineation, dehydrated during prograde metamorphism to form omphacite and garnet.

Garnets from the aluminous schists are thought by Mäkinen to show evidence of deformation in their high length:breadth ratios and general shattered, anhedral appearance (Fig. 2.11, thesis 1967). If this evidence is admitted, then the deformation affecting these garnets is very likely to be F_2 or F_3 since the garnets probably crystallized post F_1 (Atollated garnets with zoned arrangement of inclusions. Fig. 2.9. op.cit. 1967. Inclusion swirls of rutile following S_1 Fig. 2.5, op.cit. 1967). It is not altogether unreasonable that garnets from the incompetent, pelitic lithologies should suffer more from the later phases of deformation (F_2 , F_3 and F_4) than garnets in the more massive metabasic bodies.

3.6.5. Evidence for glaucophane + lawsonite paragenesis during F_1 or F_e

Felix (1972) was the first to suggest the possibility of lawsonite being stable during an early stage of the prograde metamorphic evolution of Groix. Although Mäkinen had observed rectangular and rhomb shaped pseudomorphs on Groix rather similar in appearance and paragenesis to the lawsonite porphyroblasts from Syra, Greece, he failed to conclude that these pseudomorphs were necessarily after lawsonite (Mäkinen, 1967).

It is crucial to a proper understanding of the mineral growth episodes on Groix that firstly these pseudomorphs should be positively identified as being derived from lawsonite and secondly that their relationship to F_1 or F_e deformation should be determined.

By a study of the interfacial angles on the pseudomorph forms and X-ray diffraction of the pseudomorphing minerals, Felix and Fransolet (1972) conclude that lawsonite was very probably the original mineral present. In his structuro-mineralogical study of the pseudomorphs Felix (1972) goes on to conclude the possibility of a glaucophane + lawsonite assemblage and describes the relationship of mineral foliation to pseudomorph orientation. Generally, he concludes that the pseudomorphs lie in the intense planar schistosity S_1 or S_e with the long (Z) axis of rectangular sections or the short (x) axis of rhombic (010) section parallel to the schistosity trace. In addition, he describes the preferred orientation of clinozoisite needles in the core regions as following an early schistosity " S_i " which is divergent to the schistosity he calls " S_e " defined outside the pseudomorph.

From these observations, Felix concludes that lawsonite could have been rotated whilst it was growing and that shearing movements ceased before the final stages of porphyroblast growth. In other cases, where pseudomorphs appear to be stretched and boudinaged it seems that the shearing movements that caused rotation of the porphyroblasts continued after growth ceased. An alternative interpretation, offered by the present author, is that lawsonite grew syntectonic to F_1 but helicitically over glaucophane and other minerals following an earlier F_e microfold. The fact that lawsonite pseudomorphs often are elongate in the same direction as glaucophane lineation, which is divergent to BS_1 , would suggest that lawsonite grew at the same time as the glaucophane, if not actually during F_e perhaps mimetically shortly after. This would explain the observed preferred orientation of lawsonite pseudomorphs and the occasional case of boudinage during F_1 stretching (PHOTO Fig. 10 shows a profuse development of lawsonite pseudomorphs on a S_1 surface at the margin of a metabasic boudin near the lighthouse at Pointe des Chat).



Fig. 3.10

A profuse development of pseudomorphs after lawsonite on a weathered S_1 surface at the margin of a metabasic boudin. Locality: Pointe des Chats. Compass is 10cm long.

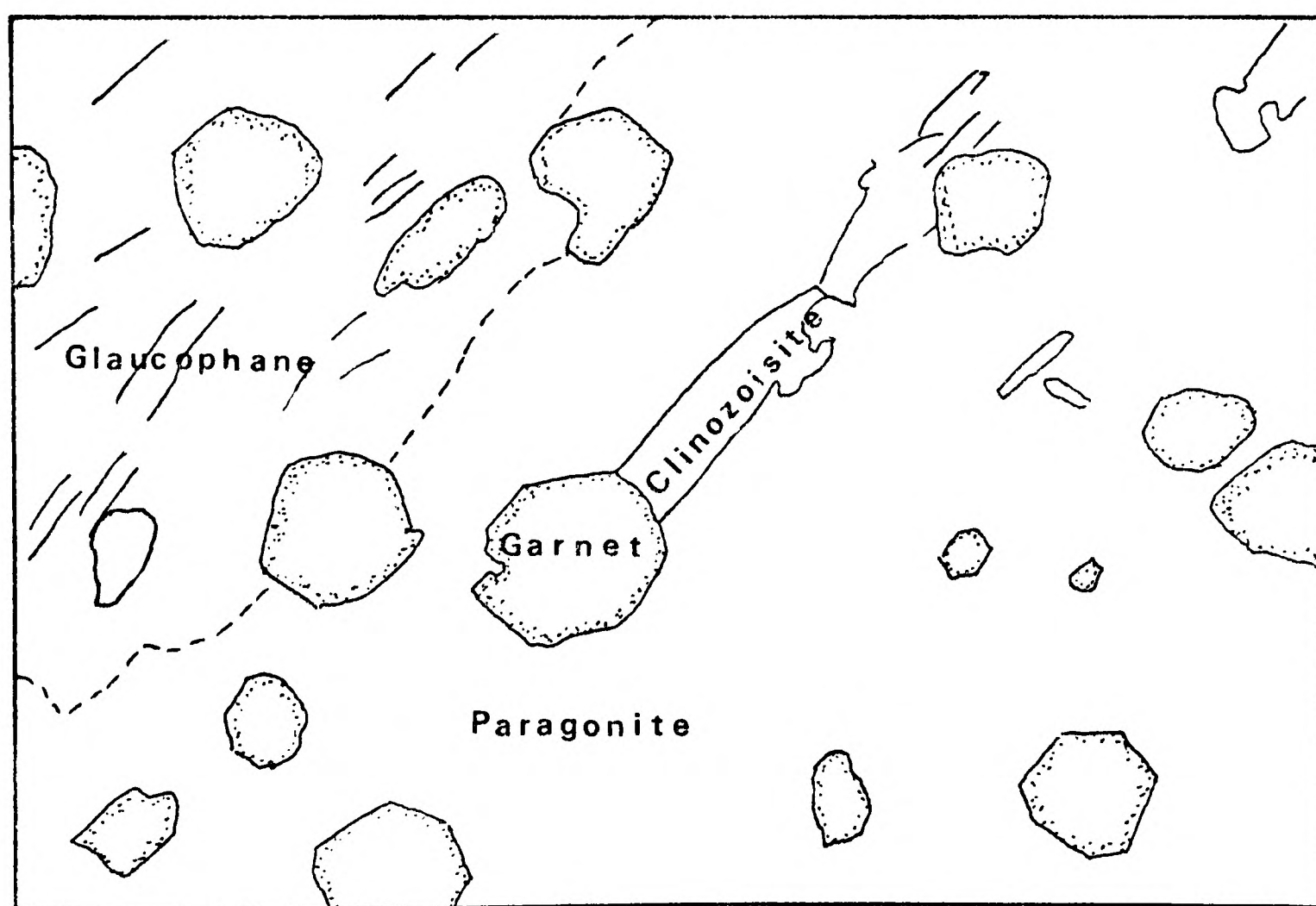
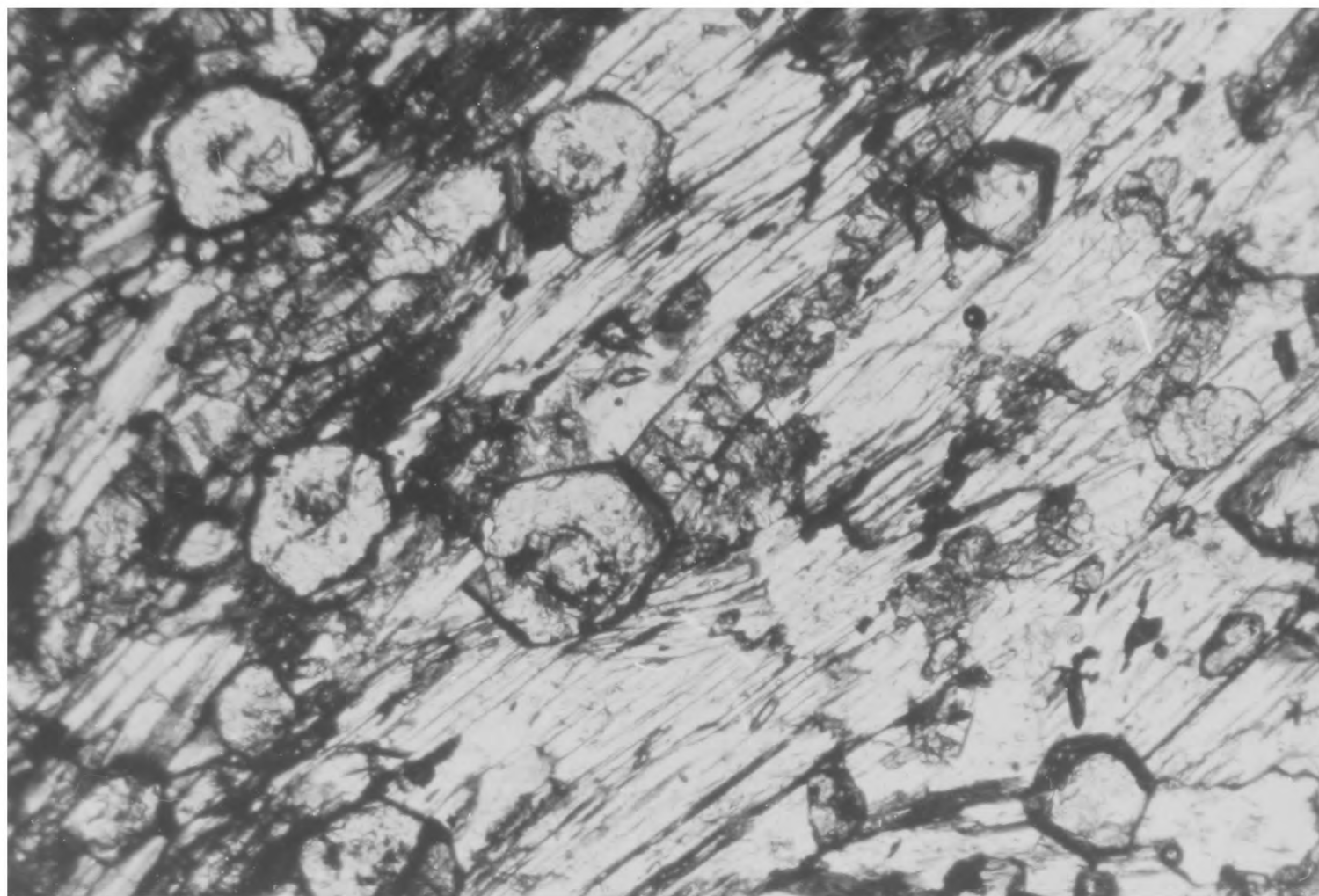
3.6.6. Augen structures do not necessarily imply pre F_1 growth of garnet omphacite assemblage

This point has been touched upon in the sub-section discussing the evidence for post tectonic garnet growth in the metabasic schists, but it is worth stressing that the occurrence of augen like structures does not demand a complex tectono-metamorphic history prior to the development of S_1 (c.f. de Wit, 1976). On the contrary, the garnet augen could easily be developed post tectonic to F_1 but mimetic to an early F_e fabric. It could be that when garnet started to grow, it nucleated preferentially parallel to S_1 and in a direction dictated by the lineation of early formed lawsonite and glaucophane. In this way, garnet augen would be formed with unstrained pyroxene laths which are elongated parallel to B_{Se} preserved in the 'pressure shadows' (see PHOTO Fig. 11 of pyroxene augen in PC10).

3.6.7. Early garnet growth is syn lawsonite breakdown but pre sphene growth in zone I metabasics

Although Felix (1972) observed small garnet euhedra within the lawsonite pseudomorphs, he failed to draw any broad conclusions about the timing of garnet growth in relation to his postulated glaucophane + lawsonite paragenesis. He concludes that, since garnet, glaucophane, epidote and sphene are found unaltered within the pseudomorphs, these phases represent original inclusions in the lawsonite before its breakdown, which have not been affected by the substitution of clinozoisite, albite and paragonite for lawsonite and glaucophane.

The present author has both observed and analysed these small (0.2 - 0.4 mm diameter) garnet euhedra in certain metabasic schists (e.g. PHOTO Fig. 12 of PM1) with pseudomorphs. It is significant that these garnets have Mn rich cores, are strongly zoned toward Fe-rich rims and contain rutile inclusions. This is most easily explained if the garnets grew



Mag. X72.

Fig. 3.12 Post-tectonic garnet euhedra in pseudomorph after lawsonite. Specimen PMI.

during prograde metamorphism as a result of lawsonite breakdown and before the development of sphene. This hypothesis is supported by the discussion of garnet forming reactions in Section 4.7, Chapter 4.

Mn-rich garnets are typically developed at the lower temperature of greenschist facies metamorphism (ca. 350-400°C, E.H. Brown, 1969) in the chlorite zone. The zoning paths, and chemical composition of garnets from Groix metabasics are very similar to those quoted by P.M. Black (1973) for the transitional and epidote zones of the Ouégoa schists, New Caledonia. A full discussion of this useful facies comparison will be given in Chapter 4 on 'the conditions of metamorphism', but it suffices here to say that the garnet zoning paths (see Appendix IV) obtained by the present author suggest that the small euhedral Mn-rich garnets (e.g. PM1) grew at an early stage of the prograde metamorphism while lawsonite was breaking down. The rims of the small garnets from pseudomorph bearing schists appear to have compositions overlapping with the cores of large, poeciloblastic garnets from omphacite and sphene bearing glaucophane schists. From this one may conclude that zoning paths in Groix garnets are caused by prograde growth, and that the large garnets grew at a stage when lawsonite had broken down and after sphene and omphacite had begun to form.

3.7. Evidence for partial and complete retrogression post F_2 deformation phase

Included in this section is a discussion of all those mineral replacing reactions and textural changes that have affected the prograde metamorphic sequence as defined in the previous sections. Generally, retrogressive phenomena on Groix can be fitted into two categories. The one, which involves only partial recrystallization of the prograde assemblage, leads to reaction rims and only very local attainment of chemical equilibrium. The other category involves a complete recryst-

alization in the greenschist facies and obliteration of all previous fabric elements.

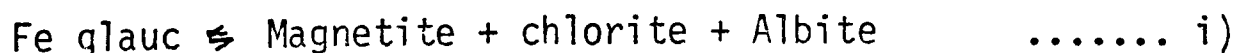
Since the degree and character of retrogression seems totally unrelated to pre-existing structure or metamorphic grade, one can assume that the retrogression proceeded everywhere under the same condition of pressure and temperature but the completeness of the chemical equilibrium attained was controlled by the availability of water. There would be differences in the reaction path towards equilibrium, according to whatever primary phases were present, but everywhere the retrogressive paths were convergent on the same P , T and X_{H_2O} . Hence, for a given lithology the maximum phase equilibrium assemblage is always the same for totally retrogressed rocks of that chemistry throughout Groix.

3.7.1. Partial retrogression

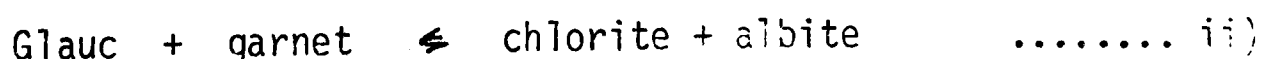
a) Pseudomorphs after glaucophane

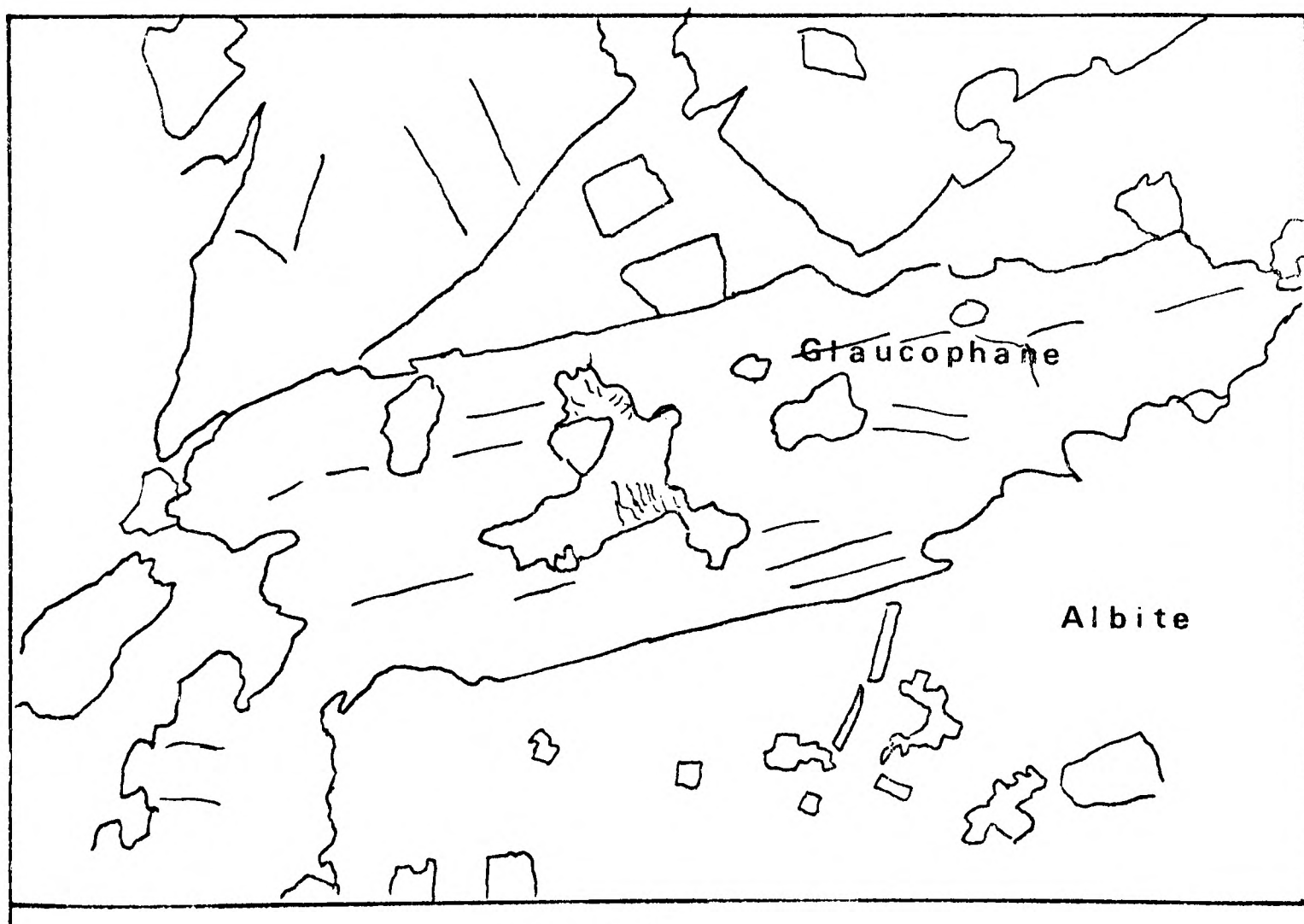
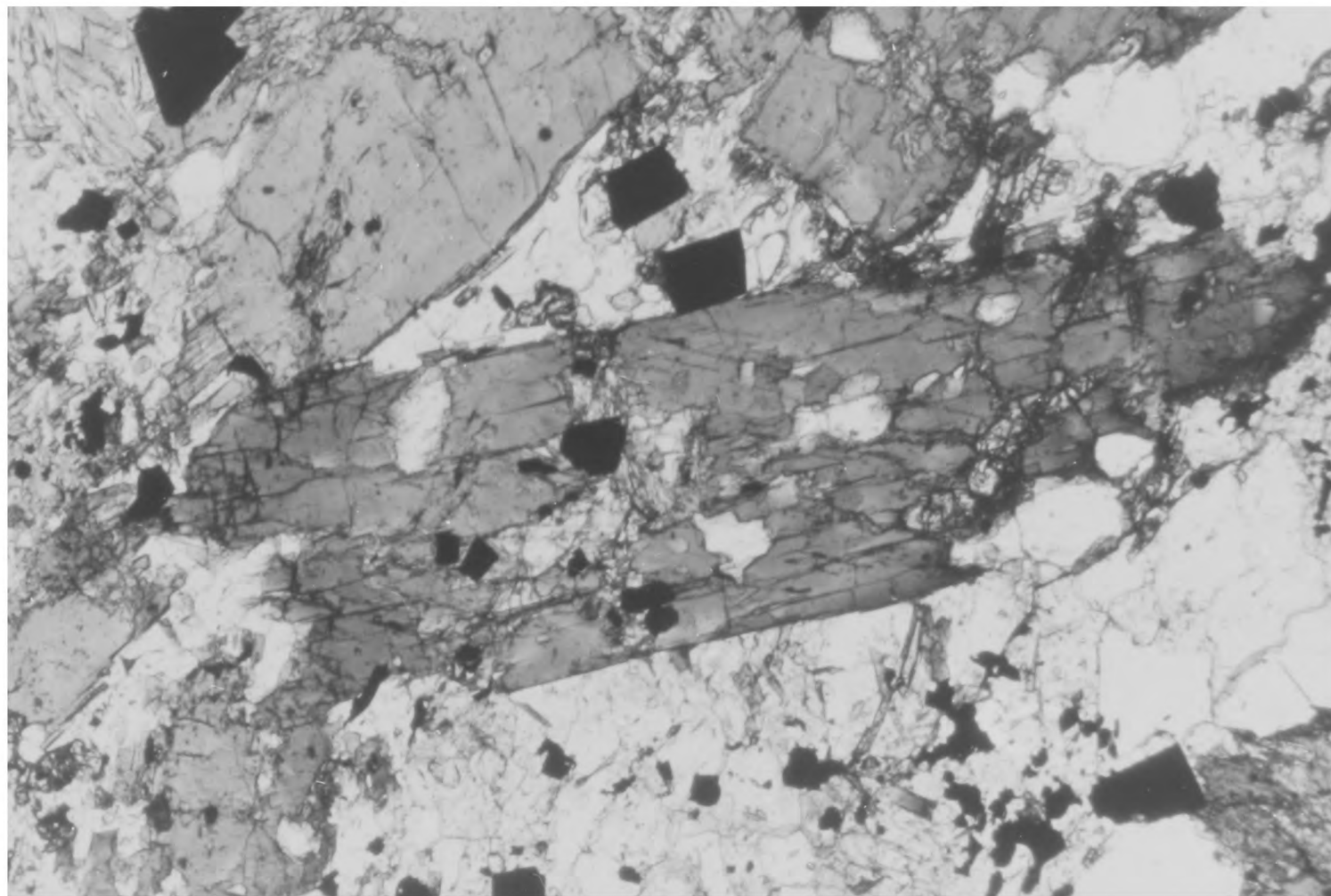
As mentioned in Section 1A of Chapter 2, replacement of glaucophane by an aggregate of chlorite, magnetite and albite is commonly observed in both garnet and non-garnet bearing varieties of the aluminous glaucophane schists.

In the case of non-garnet bearing aluminous schists and some metabasic schists without a second Ca-bearing phase (i.e. present author's KBI PHOTO Fig. 13), the glaucophane is typically Fe-rich, and the breakdown reaction may be expressed:-



With garnet-bearing pelitic assemblages it is probable that the breakdown of glaucophane proceeded by a different reaction path, and other phases in the rock may be involved.





Mag. X45.

Fig. 3.13 Alteration of ferroglaucophane to magnetite (opaque) albite, quartz and chlorite in specimen KBI.

However, ii) will only balance if the $\frac{\text{Fe}^{2+} + \text{Mg}^{2+}}{\text{Fe}^{3+} + \text{Al}^{3+}}$ ratio of the garnet involved is greater than that of the chlorite produced and if the system is Ca free. Since garnets from the aluminous schists contain up to 6% mole grossular component (c.f. LM8 Garnet analysis of Makanjuola and Howie, 1972) it is difficult to see how a univariant reaction can be written that involves only the phases seen in this rock type to produce chlorite (and possibly albite) at the expense of glaucophane.

Pseudomorphs of chlorite and albite after glaucophane have been observed in highly retrograded glaucophane schists of both pelitic and basic parentage (Rocks 97862.1, 97862.2 of the Harker Collection, 97857, 97861.1 Cambridge)

The fact that these pseudomorphs, sometimes up to 5 mm across, are contained in augen structures suggests that the retrogression of the glaucophanes pre-dated F_3 deformation (Fig. 2.1a Chapter II, Makanjuola, 1967).

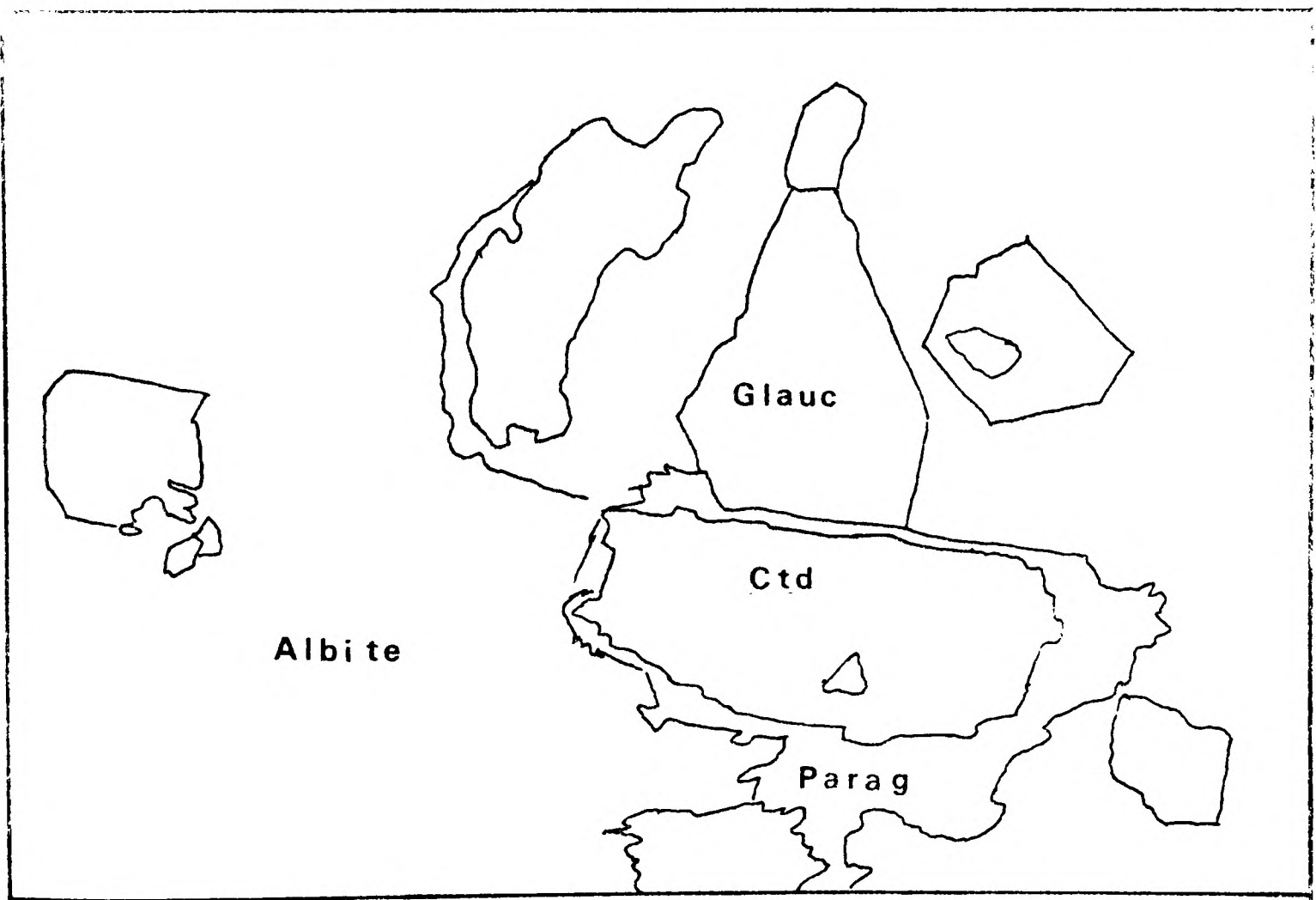
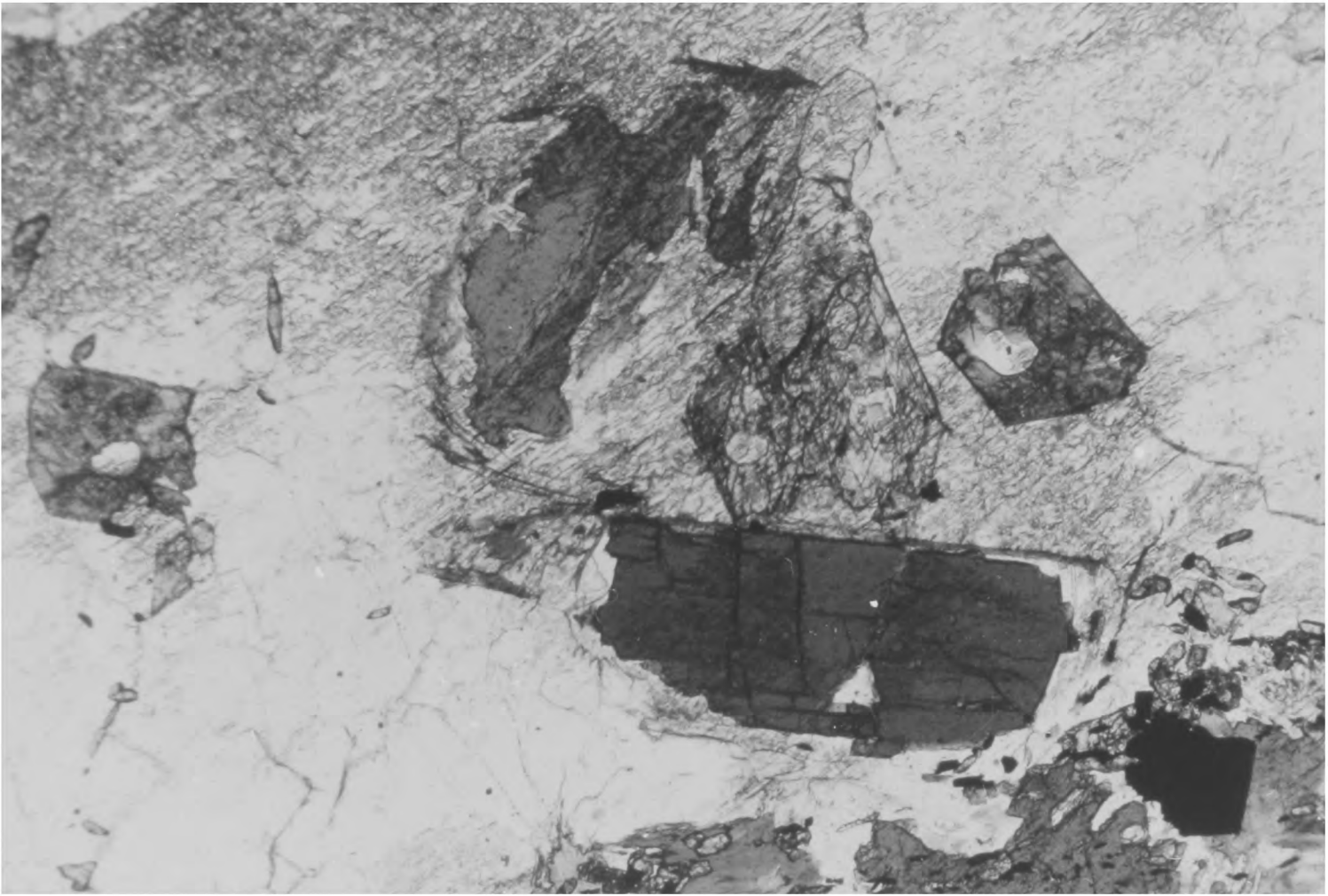
b) Chloritization of Chloritoid

The sericitic variety of chloritoid, as commonly developed in the low grade zone I phyllites, often shows alteration to chlorite and possibly stilpnomelane or green biotite (see Section 3.5(b)). Important alteration is confined to glaucophane bearing varieties (type IAB, Spec PM10) where both paragonite and chlorite are seen to rim the chloritoid porphyroblasts (see PHOTO Fig. 14).

c) Chlorite rims to garnets

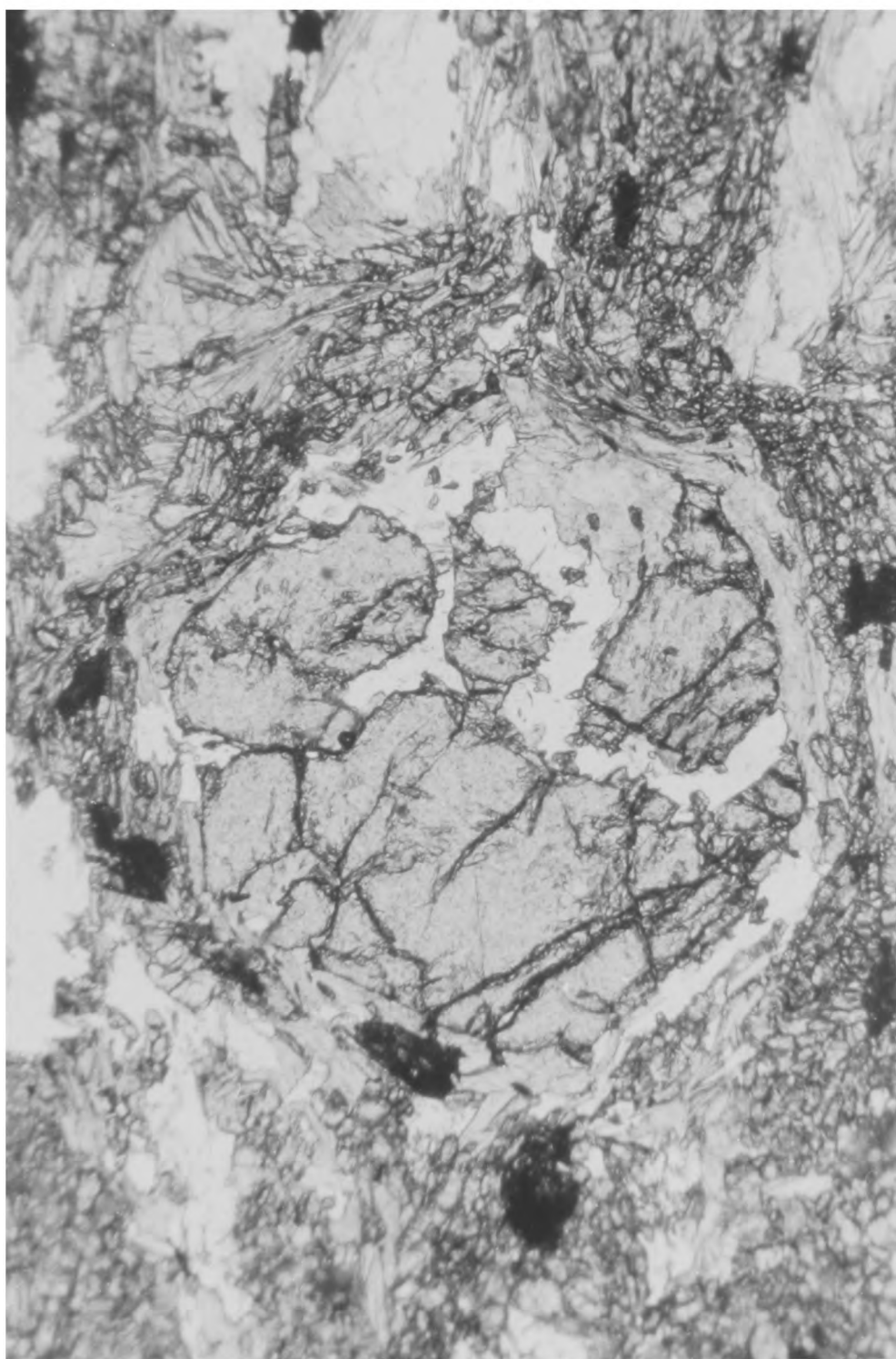
There is abundant evidence for the partial retrogression of garnets post F_2 but pre F_3 since chlorite rims are deformed by kink bands (Rock No. 97860, Harker Collection. PCl spec. of present author, see PHOTO Fig. 15).

Garnet growth in both pelitic and basic assemblages has been assigned



Mag. X45.

Fig. 3.14 Primary chloritoid (ctd) and glaucophane (glauc) altered to paragonite (parag) and albite in specimen PMIØ.



Mag. x 45

Fig. 3.15

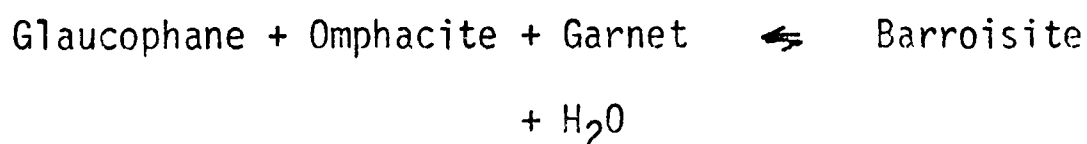
Garnet porphyroblast altered along cracks and rim to chlorite and albite. Matrix consists of clinozoisite, chlorite, albite and magnetite.

Specimen PC1.

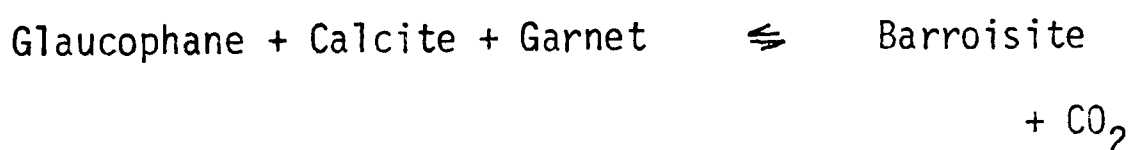
to a post F_2 episode (see Section 3.6.4) so any deformation which affects the reaction rim must by definition be F_3 . On the mesoscopic scale, F_3 is important in producing kink bands that trend E-W. (Some chlorite rims are streaked out into augen following S_1 or S_2 , but this does not mean that retrogression was necessarily syn F_1 or F_2 since the chlorite could have grown mimetically over pre-existing fabric elements.

d) Barroisite rims to glaucophane

One of the most important effects of partial retrogression in the omphacite or calcite bearing metabasic schists was the rimming of glaucophane with a blue-green sub-calcic sodic amphibole (barroisite of Binns, 1967). This interpretation is made possible because glaucophane cores are always separated from barroisite rims by a distinct chemical discontinuity (see PC5 Glaucophane and Barroisite analyses in Appendix III). Also, one can write a balanced retrogressive reaction;



which accounts for the local equilibrium achieved during water-deficient retrogression of pyroxene-bearing metabasics. For calcite bearing epidote banded schists (e.g. present author's TUD 5), the following reaction would account for the development of barroisite rims:

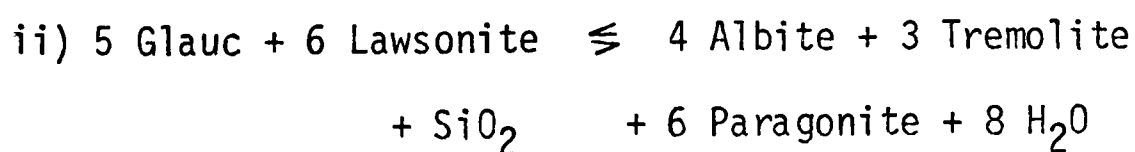
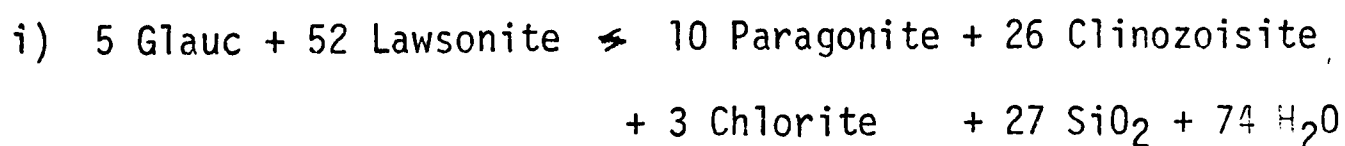


Finally, barroisite-rich separates yield K-Ar ages significantly younger than those obtained on prograde phengites (Carpenter and Civetta, 1976). This would suggest a separate growth episode for barroisite rather than an equilibrium co-existence of glaucophane/barroisite due to unmixing about a solvus in the solid solution.

e) Lawsonite breakdown in non-garnet bearing metabasics

It is apparent from a discussion of the paragenetic and chemical evidence (see Section 3.6.5) that lawsonite was stable in certain zone I (non-garnet bearing) metabasics throughout prograde metamorphism and did not break down into pseudomorph aggregates until syn F_3 retrogression. This conclusion is derived mainly from the observation of albite-rich pseudomorphs in ferro-glaucophane bearing epidote banded schists (non-garnet, non-calcite bearing, e.g. KBl) and rare glaucophane-chloritoid schists of zone I (e.g. present author's MAR 1, Makanjuola's PM6). These pseudomorphs differ from those of zone II, in their lack of garnet and a clinozoisite rich central region, with residual glaucophane. Instead the pseudomorphs of zone I contain predominant poeciloblastic albite with inclusions of white mica, chlorite and barroisite.

The present author has presented evidence to show that garnet and sphene growth in zone II is probably post-lawsonite breakdown (sections 3.6.4 and 3.6.7) so we are left with only glaucophane, and possibly epidote, as residual pre-lawsonite inclusions. This being so, then the absence of glaucophane, and presence of chlorite, albite or barroisite (which have been discussed as alteration products of glaucophane in Section 3.7.1 a)) in the pseudomorphs of zone I would demand that the pseudomorphs grew post- or syn-glaucophane breakdown and during retrogression.



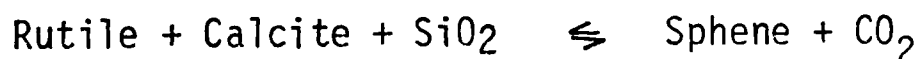
Reaction ii) is more likely in view of the fact that the Tremolite molecule could be accommodated in barroisite and that less exchange

of H_2O and SiO_2 are involved as mobile components. The actual reaction path that operated may be complicated by the presence of magnetite or calcite as buffers for the fluid phase (see Chapter 4).

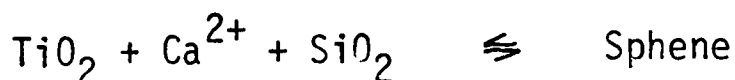
f) Sphene rims on rutile in zone I

Although the presence of primary sphene (i.e. syn prograde metamorphic growth) is a possibility in certain non-garnet bearing basic schists with abnormally low $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios (e.g. QM3 of Mäkanjuola, KBl of present author), the vast majority of epidote banded zone I schists contain rutile as inclusions with occasional secondary overgrowths of sphene. Since these rutile inclusions follow trails oblique to S_1 and porphyroblast margins, they are thought by the present author to have grown at a very early stage of the deformational history (pre F1). Sphene could have grown at any time after that, but it is considered most likely by the author that the replacement occurred during retrogression.

This is because the formation of sphene from primary rutile would require either an input of heat or a drop in p_{CO_2} during metamorphism/retrogression.



If calcite were not present, then the activity of Ca^{2+} in a Ca-bearing phase and temperature would be the controlling factors in the formation of sphene.



Low Temp
High Pressure

High Temp
Low Pressure

In the context of Zone I metamorphism, where glaucophane + rutile + lawsonite/epidote assemblages were stable at high P_{LOAD} /low temperature under variable X_{CO_2} fluid (see Chapter 4) it is more likely that sphene grew later when the temperature went up, and P_{LOAD} or X_{CO_2} went down

during retrogression.

It appears sphene could have grown in zone II after lawsonite breakdown, but prior to omphacite growth, when temperatures increased and P_{fluids} decreased during prograde metamorphism.

3.7.2. Complete retrogression of metabasic schists

Most previous authors writing about the metamorphic history of Groix recognise the importance of a late greenschist facies retrogression which has partially overprinted pre-existing assemblages.

Triboulet (1974) states that "...this second, retromorphic episode has partially transformed these rocks and has had a tendency to proliferate (at the expense of primary assemblages) albite, chlorite, epidote and blue-green amphibole in the basic rocks".

He goes on to compare the pressure/temperature conditions of the retrogressive metamorphism with those of the primary, prograde metamorphism.

The conclusion that he draws from a consideration of the constraints imposed by the prograde assemblages, and the absence of omphacite or glaucophane in the secondary assemblages, is that retrogression took place at the same temperatures as prograde metamorphism (470°C) but at lower pressures (ca. 6.5 Kbars).

The present author considers that retrogression on Groix took place at temperatures in excess of those during prograde metamorphism and to an extent controlled by the availability of water.

Where water was totally available (i.e. $P_{\text{FLUID}} = P_{\text{LOAD}}$ during retrogression) a complete recrystallization of the rock mass appears to have occurred by a process analogous to a combination of the partial retrogressions a), c), d) and e) discussed above.

The end-product of this process amongst the metabasic schists is a massive chlorite albite greenstone with accessory clinozoisite, sphene and aluminous actinolite. The lack of magnetite in these schists would

suggest that recrystallization completely re-distributed any Fe incorporated in early formed (i.e. pre or syn F_3) magnetite derived from the breakdown of glaucophane.

Similarly, the lack of rutile or ilmenite inclusion trails and the occurrence of sphene points to an obliteration of early formed fabric during retrogression and the growth of sphene under suitable conditions in the greenschist facies (see Section 3.7.1 f)).

The most characteristic textural feature of the massive greenstones is the occurrence of large (0.5 cm)gefüllte albite poeciloblasts which contain inclusions of sphene, clinozoisite and actinolite. Since the albites are unzoned, unstrained and rarely exhibit any preferred orientation they are thought by the present author to have crystallised post F_3 deformation, and to have grown over minerals formed early in the retrogression which follow S_3 .

Chlorite is generally excluded from the albite, which may indicate that chlorite was continuing to crystallize after the albite attained large size, and that its preferred orientation may be no more than mimetic over earlier fabrics.

Further evidence that complete retrogression was probably post F_3 deformation can be found in the field relations of chlorite + albite and albite + quartz vein pegmatites. These bodies, which can be found as irregularly orientated pods and stringers within unretrogressed basic horizons and the more schistose, inhomogeneous and tectonically deformed aluminous schists, cut across all F_2 open folds and F_3 kink bands.

Brière (1920) regarded the formation of albite in the schists and in the veins as contemporary events, which she described as metamorphism by impregnation. Thus, if Brière was correct and if the veins cut F_3 structures, then complete retrogression and the growth of albite poeciloblasts must be post F_3 deformation.

TABLE 3

ZONE II

ZONE I

	BASIC	ALUMINOUS	BASIC	ALUMINOUS
pre F ₁ or Fe	Rutile Granoblastic epidote Small lined glaucophanes ? Lawsonite	Rutile/ilmenite Sericitic phengite Fe-rich chloritoid Primary Fe-Al rich chlorites	Fine grained glaucophane ? Lawsonite, rutile, garnet cores Albite + clinozoisite + paragonitic pseudomorphs Sphene Pyroxene augen, glaucophane Epidote, phengite Garnet rims	Rutile/ilmenite Sericitic phengite
F ₁				
post				Mg-rich chloritoid Garnet
syn	Large, decussate glaucophanes Porphyroblastic epidote	? stilpnomelane in graphitic mica-schists		Porphyroblastic phengite
F ₂				
post		RETROGRESSION		? Fe biotite
syn	Magnetite + albite + chlorite replaces glaucophane Albite + paragonite + clinozoisite replaces lawsonite Sphene rims rutile	Chloritization of chloritoid ? stilpnomelane Sphene rims rutile	Chlorite rims to garnet Barroisite rims to glaucophane	Chloritization of chloritoid Chlorite rims to glaucophane ? stilpnomelane
F ₃				
post	Albite (gefülte) chlorite, sphene, clinozoisite, actinolite		Albite (gefülte) chlorite	
F ₄ post	Talc, actinolite, chlorite in magnesian schists	Tourmaline		Tourmaline

CHAPTER 4

CONDITIONS OF METAMORPHISM

4.1. Introduction

The problem posed in this chapter is to infer, from detailed chemical analyses of co-existing minerals, the metamorphic conditions that operated during the formation of the mineral assemblages described in Chapter 2.

First of all, it is necessary to establish, or at least assume, that certain mineral assemblages observed in rocks of given bulk chemistry represent the attainment of thermodynamic equilibrium in that system. Chemical equilibrium, as defined in classical thermodynamics, can be said to have occurred in a polyphase, multicomponent system when the chemical potential (μ_i) of component i is the same in all phases of the system at the pressure and temperature of interest.

$$\mu_i = \mu_i^0 + RT \ln a_i$$

where μ_i^0 is the chemical potential of component i at an arbitrary standard state (in this case, at $T = 298^{\circ}\text{K}$ and P of interest), R is the gas constant, T the equilibrium temperature and a_i is the activity of component i (usually a simple function of concentration). This state can only be achieved if the system is held at constant pressure, temperature and composition for long enough so that exchange reactions between co-existing phases can reach equilibrium.

Thermodynamics can predict the probability of certain reactions occurring on the basis of known thermochemical parameters (Entropy, enthalpy and Gibb's Free energy of participating phases) or experimentally determined equilibria. But kinetic factors decide whether or not predicted reactions will take place, and are thus important in controlling the development of equilibrium assemblages and textures in

metamorphic rocks.

Even at low temperatures ($< 300^{\circ}\text{C}$) where reaction rates are slow by comparison with natural mineral growth and strain rates, equilibrium can be achieved if there is a continuous re-crystallization and mixing of reactants throughout prograde metamorphism. In this case, the conditions of metamorphism inferred from the observed mineral assemblage would refer to the peak (highest temperature) attained, when reaction rates were at their fastest, provided that quenching of the equilibrium was fast enough during uplift. Taking these comments, and the opinions of many metamorphic petrologists (including Vernon, 1976, Spry, 1969), into account, one can arrive at a list of criteria for the stable co-existence of minerals.

1. The number of phases in equilibrium must not exceed by more than two the number of components needed to define those phases, i.e.

$$F = C + 2 - P \quad \text{Gibb's Phase Rule}$$

where P = maximum number of phases at equilibrium.

C = minimum number of independent chemical components needed to describe phases.

F = degree of freedom of the system, or variance.

In a multicomponent system, such as natural silicate mixtures, one can have a situation of divariance ($F = 2$) where P , T are free to vary independently of each other with no change in the number of phases present. If one extra phase is added, then P and T are constrained to vary dependently, and $F = 1$.

At an invariant point in P, T space occurs the unique co-existence of the maximum number of phases, and any change in conditions will cause the loss of one or more phases.

Thus, once we have chosen suitable components for a system, it is possible to say whether a particular assemblage is over- or under-defined in terms of variance.

If a system has more phases than can possibly co-exist at the invariant point, then it is said to be over-defined and either one is dealing with a disequilibrium assemblage, or a component whose addition does not involve an extra phase (i.e. $X_{H_2O} < 1.0$ due to presence of CO_2 , CH_4 etc.).

Most metamorphic rocks are underdefined in terms of the number of possible components, and show rather restricted mineral assemblages ($P < 6$). This would imply that metamorphic mineral assemblages define divariant fields in P,T space (corresponding with facies) and that any further attempt to decrease variance must rely on determination of the activities of end-members in the solid solutions that make up the constituent phases. This last remark refers particularly to exchange equilibria between end-member species in homogeneous phases, i.e.

Diopside - Hedenbergite - Jadeite solid solution in Omphacites
Almandine - Pyrope - Grossular solid solution in Garnets)

So, in summary, it is fair to say that the remarkably small number of phases and the relative homogeneity of the co-existing phases argue for the attainment of equilibrium in metamorphic rocks of a given facies.

2. All phases of an equilibrium assemblage should be in contact with one another, so that no single phase is isolated from reaction by being included. In a domain where chemical equilibrium has been achieved by the diffusion of reacting species across grain boundaries during mineral growth, each participating phase must be in contact with $P-1$ other phases. Although equilibrium is very unlikely to be achieved between the separate domains of a complex polyphase system, the chances of adjacent grains within one domain reaching equilibrium are quite high. The size of the domain over which equilibrium is attained depends largely on the solid diffusion rates of exchangeable cations in crystal

lattices, and the grain size and modal abundance of the crystals concerned.

Thus, at low temperatures, where solid diffusion is slow, the domains may be very small and may include only parts of adjoining crystals. In that case, it becomes necessary to use an electron microprobe to investigate the equilibrium assemblages defined by spot analyses near grain boundaries.

In this study, microprobe analyses have been used in the assessment and calculation of equilibrium conditions rather than conventional wet chemical separate analyses (Makanjuola and Howie, 1972, Triboulet, 1974) which lack detailed positional and textural information.

3. The rim composition for different grains of the same phase should approach a constant value, and if this is true for all phases there should be a similar distribution of components between pairs of phases throughout the domain of equilibrium.

Partition or distribution coefficients for components between pairs of phases can be regarded as equilibrium constants for exchange reactions at constant pressure and temperature, where the components are defined as the end member species of the solid solution.

Thus partition coefficients can be measured directly, and should be constant throughout a domain equilibrated at one set of pressure and temperature conditions. A fuller discussion of the dependence of K on temperature and pressure is pursued in section 9 of this Chapter, which deals with the interpretation of partition coefficient data from eclogitic rocks.

4. A further test of equilibrium is provided by the fractionation of oxygen isotopes between co-existing minerals in the assemblage under consideration. Fortunately, a considerable body of data exists for

minerals in blueschist facies rocks that enables us to check oxygen isotope temperatures (e.g. Δ Quartz-Muscovite) against empirical geothermometer based on K_D values (Råheim & Green, 1974) that assume equilibrium.

If a consistent pattern of temperatures is obtained in a prograde metamorphic sequence (as observed by Taylor & Coleman, 1968, and Black, 1974) this argues very strongly for the attainment, and preservation, of equilibria corresponding to the peak of metamorphism in any one outcrop.

Conversely, any disturbance of non-attainment of equilibrium should be easily spotted by a reversal of the normal expected δO^{18} fractionation between two phases, or at least a discordance in temperatures derived from three or more co-existing minerals.

Although oxygen isotope data are not available for the mineral assemblages on Groix, they nevertheless provide some justification for the extrapolation of the Råheim and Green geothermometer to low temperatures since there is good agreement between the oxygen isotope and garnet/clinopyroxene temperatures derived from rocks of similar facies to those on Groix.

4.2. Schreinemaker's analysis of stable phase assemblages

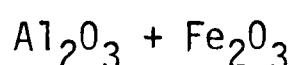
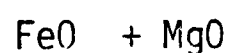
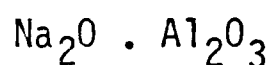
Having decided upon some criteria for recognising equilibrium in metamorphic mineral assemblages, we can now proceed to identify stable phase assemblages amongst the metabasic rocks of Groix.

In principle, one should be able to combine all the information about composition of co-existing phases for one bulk chemistry at different P and T in the form of a family of univariant equilibria derived theoretically by Schreinemaker's method (Zen, 1966). Using the Phase Rule (see section 4.1.1) a matrix of all possible reactions between discrete phases can be used to predict the topology of inter-

secting equilibria in PT space, as long as the numbers of phases and components are fixed. The end product of such a procedure is a petrogenetic grid that hopefully relates to the parageneses as seen in outcrop. Unfortunately, Schreinemaker's construction is very difficult to apply to systems where the phases show solid solution through a large range of P and T conditions, since this introduces further degrees of freedom not accounted for by the Phase Rule.

For the purposes of this study, various simplifying assumptions can be made in order to construct model systems that yield qualitative information about the more complex, natural systems.

1. Metabasic compositions can be expressed in terms of four components:



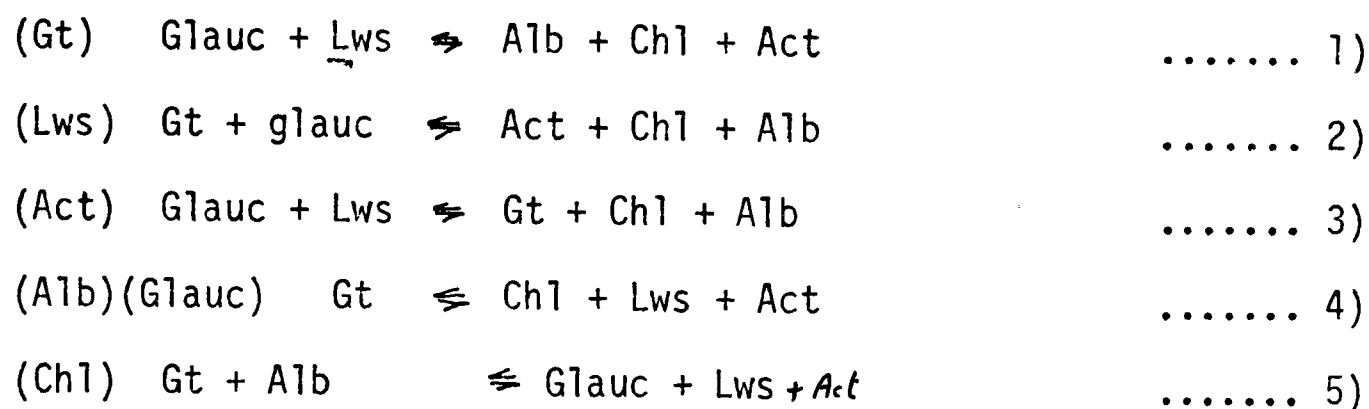
2. SiO_2 and H_2O can be counted as mobile phases, always present in excess.
3. Phases have a constant composition throughout the P,T field of interest. If it is necessary to take solid solution into account then, the end-member species of a particular solid solution cannot be regarded as separate, discrete phases in the same system since then the Phase Rule would be violated.

4.3. Example 1 (c.f. Ernst, 1963)

6 phases:	Albite	$\text{Na AlSi}_3\text{O}_8$	: Alb
	Garnet	$\text{CaFe}_2\text{Al}_2\text{Si}_3\text{O}_{12}$	: Gt
	Glaucophane	$\text{Na}_2\text{Mg}_3\text{Al}_2\text{Si}_8\text{O}_{22}(\text{OH})$	: Glauc
	Lawsonite	$\text{CaAl}_2\text{Si}_2\text{O}_7(\text{OH})_2 \cdot \text{H}_2\text{O}$	: Lws
	Actinolite	$\text{Ca}_2\text{Fe}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$	: Act
	Chlorite	$\text{Mg}_5\text{AlSi}_3\text{Al}_{10}\text{O}_{40}(\text{OH})_8$	: Chl

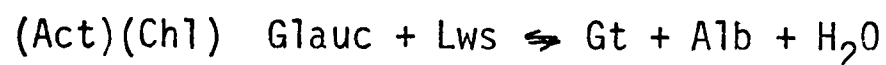
4 Components: $\text{Na}_2\text{O} \cdot \text{Al}_2\text{O}_3$, $\text{MgO} + \text{FeO}$, $\text{Fe}_2\text{O}_3 + \text{Al}_2\text{O}_3$, CaO .

There are 6 reactions about an invariant point in P,T space.



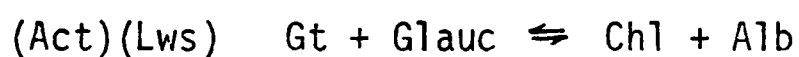
With a small amount of Ca component in the garnet, reaction 4) becomes degenerate, which means that it will have stable extensions both sides of the invariant point in P,T space.

For a unique value of Ca content in garnet, reaction 3) will become equivalent to reaction 5) and another degenerate reaction can be written:



This special case would only occur for a particular bulk composition lying at the cross-over of the Glauc-Lws and Garnet-Albite tie lines.

Another special case would occur in Ca-free compositions, whereby reaction 2) and 3) would amalgamate to form a degenerate equilibrium



This would only balance if the Al content of the chlorite were less than that of the garnet.

So, for a normal range of basic rocks lying in the Lawsonite-Glaucophane-Chlorite-Actinolite sub-tetrahedron there are five possible reactions that define stable divariant assemblages in a 6 phase, 4 component system (see Fig. 1).

After the 5 reactions were balanced, it was found that water must be evolved in all but one of the reactions, and then the assemblages with free water were arranged on the right hand side of the equations:

Fig. 4.1 Model petrogenetic grid for basic assemblages on Groix.

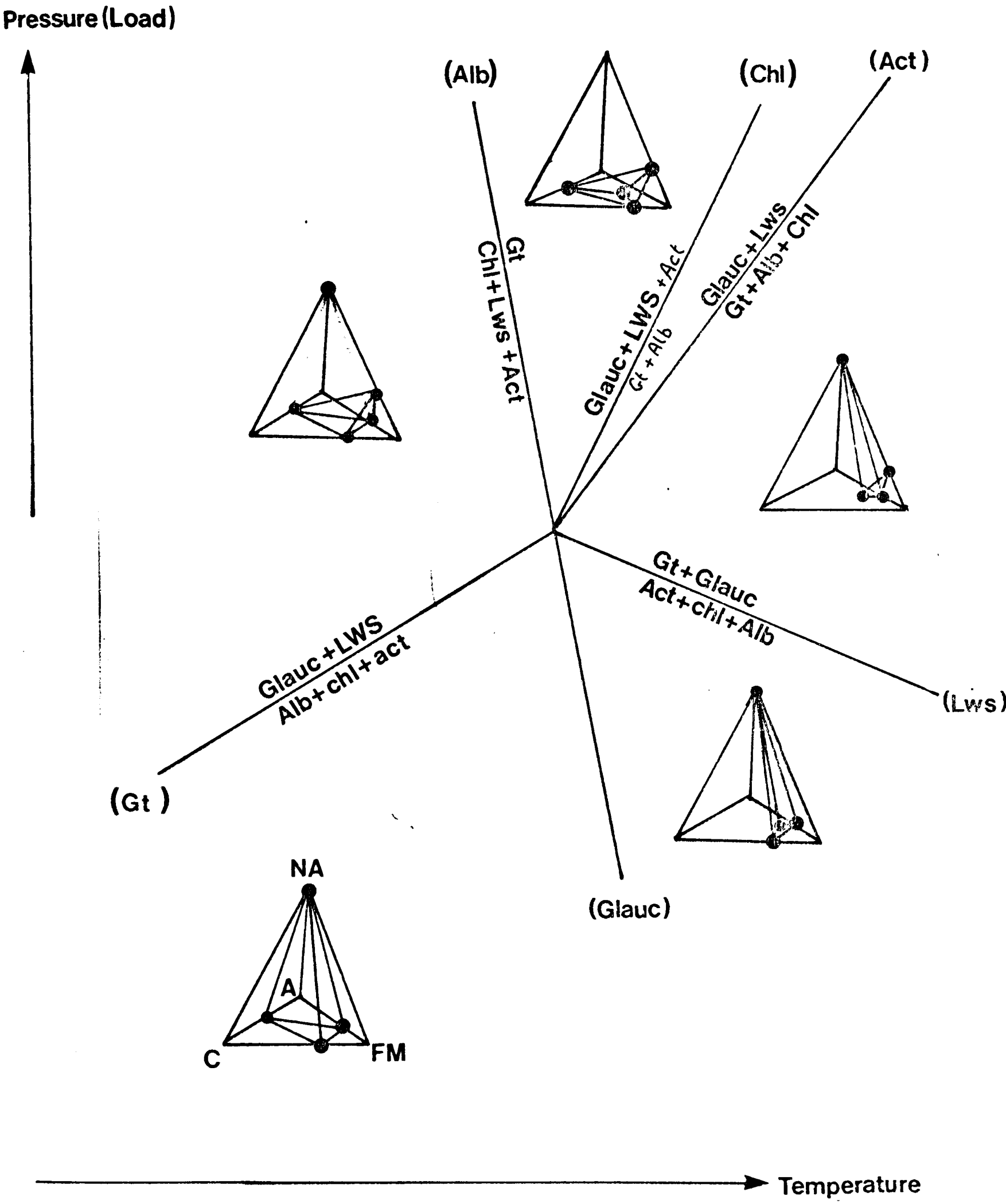
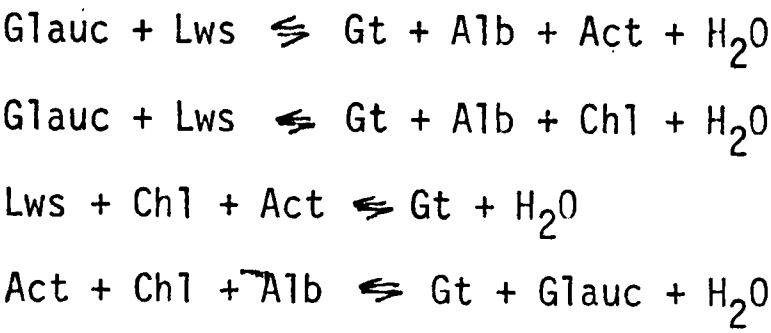


TABLE 1

Phase	298 ⁰ K, 1 bar		600 ⁰ K, 3000 bars	
	Molar volume ~ in cm ²	3rd Law Entropy in Cal/deg/mole	Molar volume	Entropy
Mg glaucophane	264	130	Same	249 ± 50 (1)
Fe glaucophane	269.47			215 ± 80 (2)
Tremolite	277.92	131.19		261.13 (3)
Actinolite	276.8	151.4		281 (4)
Mg chlorite	207.94	109		213 (5)
Fe chlorite	213.65	141.6		267 (6)
Grossular garnet	125.3	57.7		124.32 (7)
Almandine garnet	115.27	68.3		134.9 (8)
Lawsonite	101.32	56.79		117 (9)
Low Albite	100.07	50.2		97.32 (10)
Quartz	22.688	9.88	19.74	20.41 (10)
H ₂ O (liquid)	18			27.7 (11)

- (1) ΣS oxides and including volume correction
quoted in Ernst (1963).Values
- (2) Data from Hoffmann (1972); breakdown of ferroglaucophane at 633⁰K, 3kb.
Assuming dP/dT = 150 bar/⁰C.
- (3) Robie & Waldbaum (1968).
- (4) Extrapolated from tremolite data.
- (5) Deer et al. (1962). Also ΣS oxides quoted in Ernst (1963).
- (6) Data quoted in R. Powell (1973).Extracted from equilibria, based
on hercynite/magnetite entropies.
- (7) Robie & Waldbaum (1968).
- (8) Extrapolated from grossular data.
- (9) Ernst (1963).
- (10) Robie & Waldbaum. (1968)
- (11) Burnham, Holloway & Davis (1969).



Since input of heat into dehydration reactions should push the equilibrium towards the right, then the right hand side of these equations must produce the higher temperature assemblage.

It is with this last point in mind that the petrogenetic grid was constructed - those reactions which evolved most water per mole of reactants were given the steepest positive ($\frac{dP}{dT}$) slope.

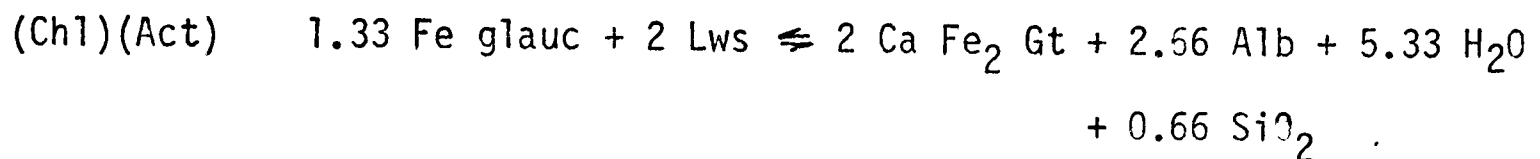
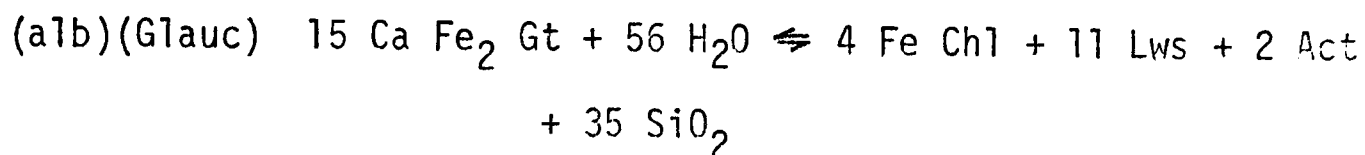
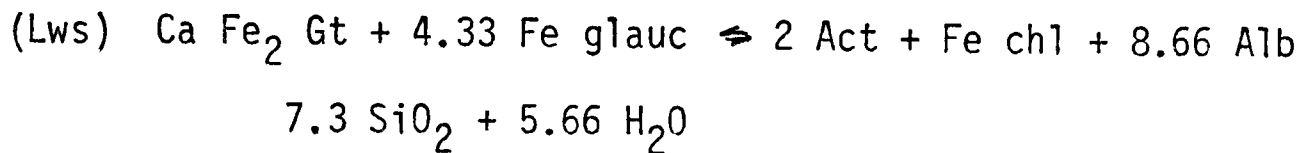
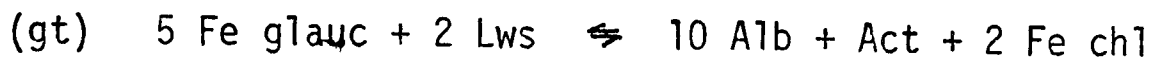
Although the slopes chosen for the reactions are purely diagrammatic and arbitrary, it was possible to calculate ($\frac{dP}{dT}$) for reactions 1) and 5) using thermodynamic data tabulated in Ernst (1963) and Robie and Waldbaum (1968), (see Table 1) and applying the Clausius Clapeyron equation.

However, there are great uncertainties in assigning molar entropies to phases when the only basis is summation of the constituent oxide entropies or extraction from experimentally determined breakdown curves (Ernst, 1961, Hoffmann, 1972).

Unfortunately, for the pure Fe end member system illustrated here (Fig. 1) these uncertainties are so large as to invalidate any conclusions drawn about the relative ($\frac{dP}{dT}$) values of reactions in the grid.

TABLE 2				
Reaction	Δ Sr cal/deg	Δ Vr cal/bar	$(dP/dT)_{600^\circ K}^{3kb}$ bars/ $^\circ C$ Fe end members (Mg)	
(Gt)	481	3.7	130 $\pm$ 40	(45)
(Chl)(Act)	162	1.46	110 $\pm$ 20	(70)
(Lws)	300	1.68	180 $\pm$ 50	(40)
(Alb)Glauc	1310	16.7	80 $\pm$ 10	(90)

The balanced reactions used in the above calculations are as follows:



By using an Mg-end member system, one more nearly approaches the observed compositions of glaucophane, chlorite and actinolite/tremolite in metabasic parageneses on Groix, whilst at the same time avoiding some of the worst sources of errors in the estimation of entropy.

The figures in brackets in Table 2 are (dP/dT) values estimated for the Mg end member system, and it can be seen that predicted reaction slopes are much less than in the Fe system. Furthermore, these predicted slopes are all positive and agree with the topology constraints of Fig. 1.

4.4. 7 phase, 4 component multisystem for metabasic parageneses on Groix

Using the topology constraints for mineral compatibilities as shown in Fig. 1, it is possible to extend the petrogenetic grid to include clinopyroxene, an important phase in dehydrated zone II assemblages. By so doing, it becomes possible to delimit the path of prograde metamorphism through the grid and define divariant or univariant equilibrium assemblages in a system of restricted bulk chemistry.

In order to construct the bundle of possible univariant equilibria in this system, it is necessary to make various simplifying assumptions about phase composition. The phases used are:

Almandine Garnet	$M_3 A S_3$
Mg Glaucophane	$N A M_3 S_8$
Mg Chlorite	$M_5 A S_3$
Actinolite	$C_2 M_5 S_8 \cdot H_2O$
Diopside	$C M S$
Albite	$\frac{1}{2}(N A S_6)$
Lawsonite	$C A S_2 \cdot 2 H_2O$

where the components are defined as:

NA	=	$Na_2O \cdot Al_2O_3$
A	=	$Al_2O_3 + Fe_2O_3$
M	=	$FeO + MgO$
C	=	CaO

with SiO_2 (S) and H_2O in excess.

In using the end-member almandine garnet, and a chlorite composition less rich in Al_2O_3 than this garnet, one avoids the problem of reactions that are degenerate for a critical value of bulk composition (viz $Alm_{87.5} Gross_{12.5} \rightleftharpoons \text{Chlorite} + \text{Lws}$)

This simplification also allows us to write the degenerate reaction:



without considering the effect of grossular solid solution in the garnet.

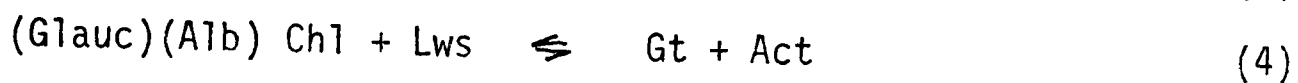
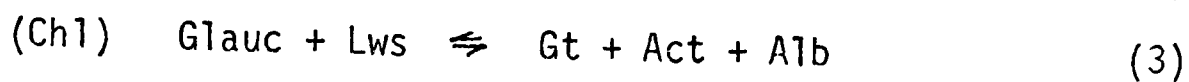
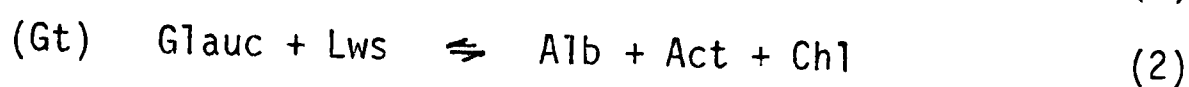
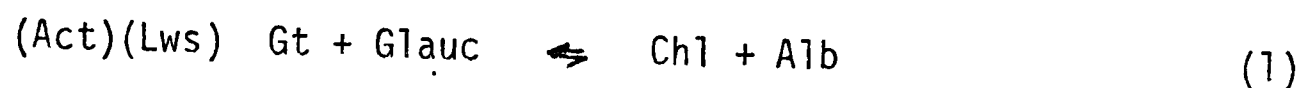
In fact, the arbitrary assumptions about phase compositions made here are particularly suitable for treating pyroxene-bearing metabasics, relatively low in CaO and Al_2O_3 component and having bulk compositions lying near the almandine-diopside tie line in C A FM space (see Fig. 3 of Chapter 2).

In the multisystem illustrated in Fig. 2, there are 6 invariant points, due to the fact that the (Alb) and (Glauc) invariant points are superimposed. Theoretically, there are 42 possible different

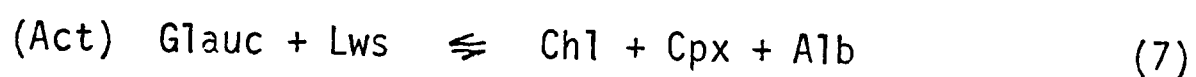
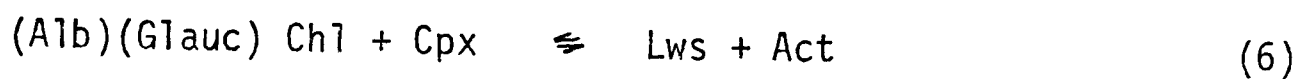
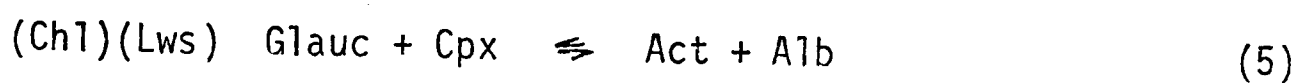
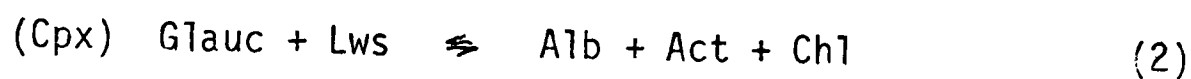
combinations of 5 phases selected from a 7 phase system, but in practice the number of reactions generated in this multisystem is reduced to 11 because of degeneracy.

Also, only those reactions defining 4 of the 6 invariants are listed below since two invariant points, (Lws) and (Act), are metastable.

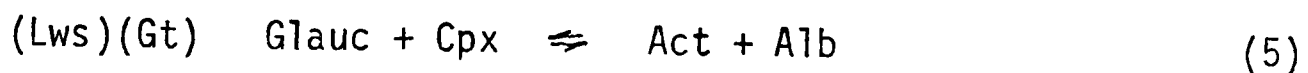
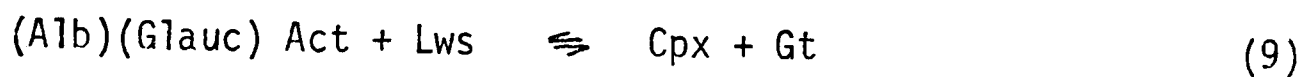
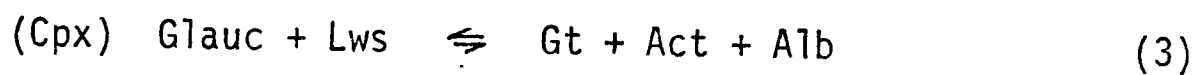
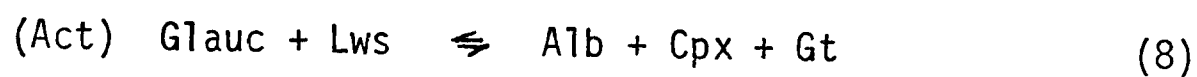
(Cpx) invariant:



(Gt) invariant:



(Chl) invariant:



(Glauc)(Alb) invariant:

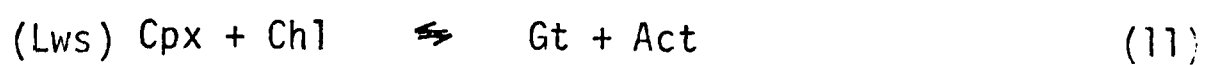
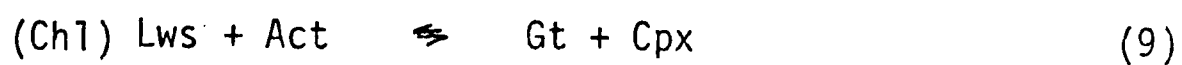
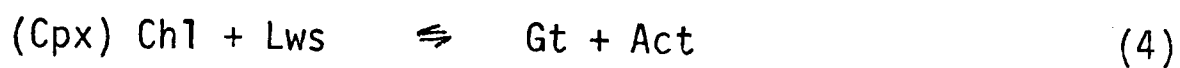
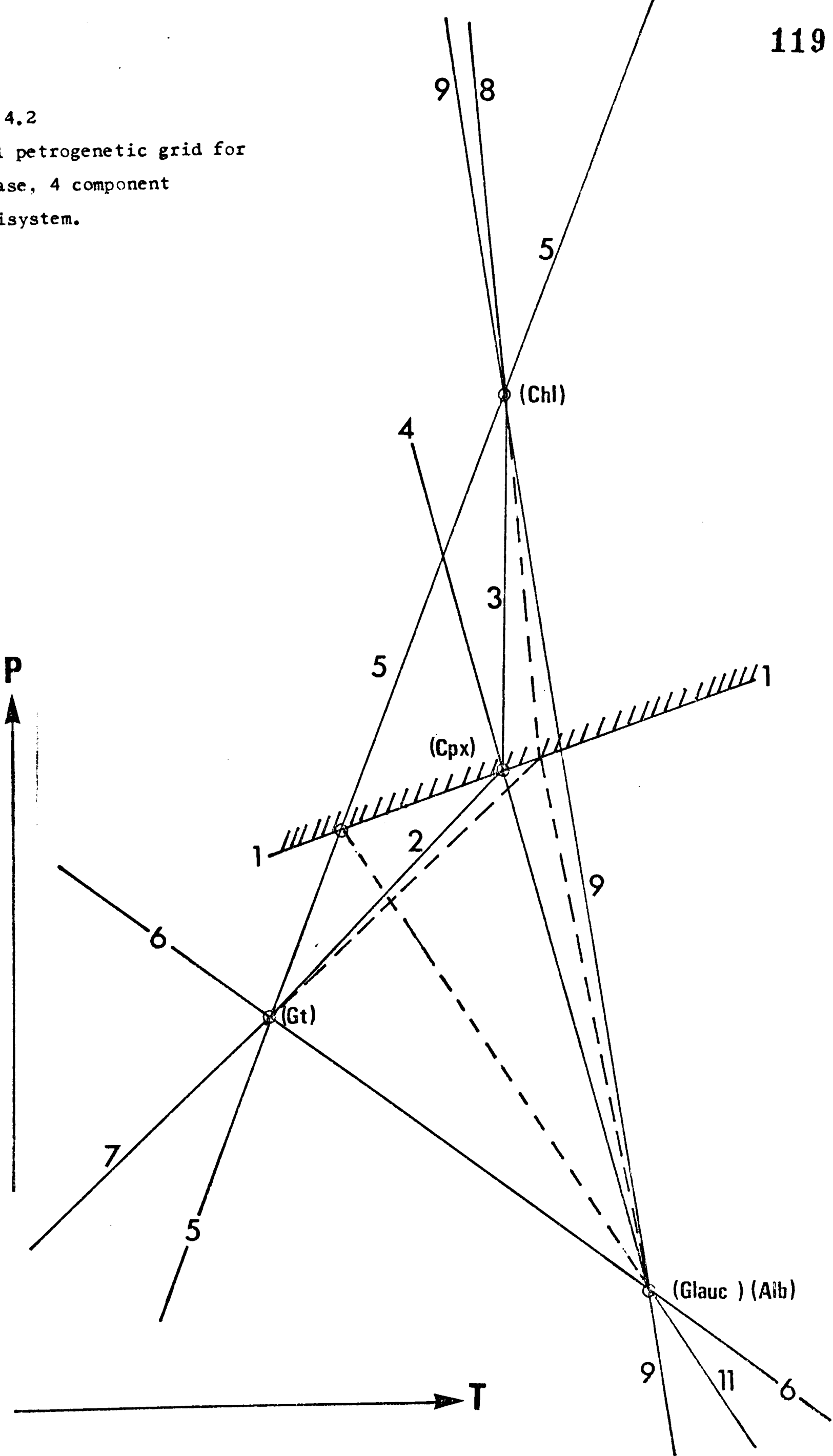


Fig. 4.2
Model petrogenetic grid for
7 phase, 4 component
multisystem.



This multisystem, or grid of invariant points, is useful in that it provides a division of P,T space into divariant equilibrium assemblage fields as follows:

Glaucophane + Lawsonite

Bounded by reactions (1), (2) and (7), and restricted to an intermediate pressure and low temperature field. This corresponds with zone I on Groix (with low fO_2).

Glaucophane + Lawsonite + Garnet

Bounded by reaction (1), which has a very shallow positive, or possibly negative slope, and reactions (3) and (8). Assemblage is stable at high pressures and low temperatures, but may be restricted to high Fe compositions, where garnet could be stabilized over a small temperature range before lawsonite breakdown. Thus, with Mg-rich compositions, it is more likely that reaction (1) is steeply negative, so the divariant assemblage field shrinks to a narrow wedge expanding with increasing pressure.

Garnet + Clinopyroxene + Albite $\pm$ Lawsonite

A high temperature, high pressure stability field, bounded by reactions (9) and (5), and invariant point (Ch1) at the low temperature end. At these pressures and temperatures in the natural system, albite would dissolve into the diopside to form omphacitic pyroxene, and an eclogite assemblage would result. To the low temperature side of (5), glaucophane can co-exist with the garnet and clinopyroxene. Over a small part of the range, between reactions (9) and (8), lawsonite is in equilibrium at low fO_2 . This facies is an analogue of the glaucophane eclogites of Groix.

Garnet + Glaucophane

A field in which glaucophane is stable outside the lawsonite-albite facies (Winkler, 1967). However, the size of the field is critically dependent on the slope chosen for reaction (1), which forms the low pressure boundary. If (1) is moderately positive, as seems possible for Mg rich compositions, then the field expands as a wedge towards higher pressures and temperatures. With a steeply negative slope, however, the stability field would expand into regions of low pressure -a result which is considered unlikely in view of the topology constraints imposed by the grid.

Garnet + Chlorite + Albite

An intermediate pressure, high temperature facies bounded by reaction (4) on the low temperature side and reaction (1) at higher pressures. This facies broadly corresponds with the garnet-bearing greenschists (e.g. Ryoke belt of Japan) and any up temperature transitions indicated on the petrogenetic grid cannot be considered realistic. This is because the model system chosen is only relevant over the P and T and fO_2 range where the 7 phases are intrinsically stable, and any attempt to extend it over all P,T space is invalid.

Albite + Chlorite + Actinolite

A low temperature, low pressure facies bounded by reactions (5), (2) and (4), and corresponding with the greenschists.

4.5. Comparison of model petrogenetic grid relations with natural metabasic assemblages on Groix

On the basis of the model petrogenetic grids presented in Figs. 1 and 2, it is possible to make predictions about the relative pressures and temperatures of equilibration for observed parageneses.

In addition, one can use the grid to identify important facies boundaries that may correspond to isograds in the field, and thus test the path of prograde metamorphism as postulated in Chapters 2 and 3 (see Table 2, Chapter 3).

However, various difficulties arise when one attempts to apply the conclusions drawn from a theoretical approach, where simplifying assumptions are justified, to a system where bulk and phase compositions are complexly variable.

The natural phase assemblages differ from model assemblages in the following important ways.

1. Chlorites from metabasics on Groix have compositions which closely reflect the bulk rock chemistry, but with a tendency to more Mg-rich composition with increasing grade. This particular solid-solution behaviour does not affect the projection point in NA C FM A space, but Al content does. The present author has found chlorites from glaucophane + epidote + chlorite schists to have compositions approaching $FM : A = 3 : 1 ((Fe^{2+}, Mg)_{4.4} (Al^{VI}, Fe^{3+})_{1.5} Al^{IV}_{1.275} Si_{2.725})$ rather than the $FM : A = 5 : 1 ((Fe^{2+}, Mg)_5 (Al^{VI}, Fe^{3+})_{1.0} Al^{IV}_1 Si_3)$ assumed for the purposes of the petrogenetic grid. This means that the projection point for the metabasic chlorites from Groix lies in almost exactly the same position as almandine-pyrophe garnet ($FM : A = 3 : 1$), and consequently the reaction coefficients for balancing possible equilibria are changed.

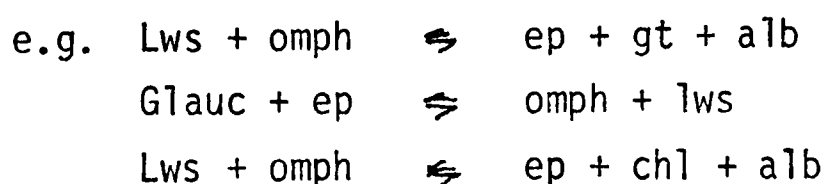
In particular, reaction (Lws) of Fig. 1 would be given a shallower, and possibly negative, slope in P,T space if the equilibrium were

balanced for more aluminous chlorite. The same effect applies to the slope of reaction (1) in Fig. 2, but the use of a high aluminous chlorite in this system is not compatible with the choice of almandine-pyropo end member garnet and the omission of actinolite as a phase.

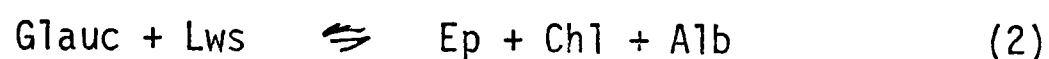
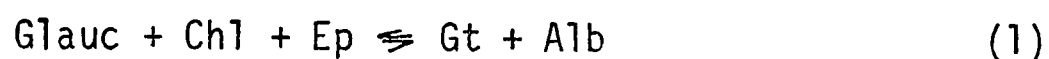
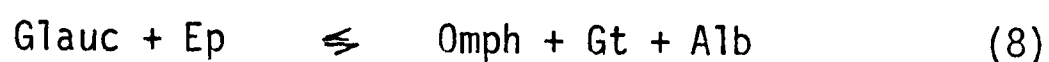
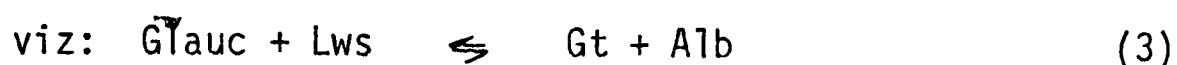
2. Both lawsonite and epidote would have to be included in the natural phase assemblages on Groix. Since the occurrence of either or both of these phases can be controlled by fO_2 independently of P or T, the topology of a P,T Schreinemaker's grid alone cannot be expected to provide meaningful predictions about stability.

3. Garnets from metabasic schists vary in composition, according to grade and bulk rock chemistry, from 15 - 25 mole percent of grossular end-member. Also, they exhibit marked zoning (see Appendix IV), which probably reflects changing P,T conditions during growth. For these reasons, garnet-forming reactions on a P,T grid are no longer univariant, and must be considered as sliding reactions.

4. Clinopyroxenes on Groix have been classified as chloromelanites (Makanjuola and Howie, 1972) and contain 28 - 35 mole percent jadeite end-member. The effect of a NA component in the pyroxene is to preclude the possibility of a (glauc) and (alb) invariant lying at the same point in P,T space. Also, the possibility arises of whole series of reactions critically dependent on whole rock compositions that can only take place when the rock falls outside the Glaucophane-Lawsonite-Chlorite-Actinolite (Fig. 1) or the Glaucophane-Lawsonite-Chlorite-Diopside (Fig. 2) subtetrahedrons.



5. For normal basic assemblages, relatively few realistic reactions can be written between all the relevant phases in the natural system (see Fig. 3).



These reactions are arrived at by considering only those rock compositions plotting near PC10, LM12, LIN-1 in the NA C FM A tetrahedron.

4.6. Conclusions drawn from comparison of model petrogenetic grids and natural metabasic assemblages

Despite all the problems that arise in the use of Schreinemaker's construction to imitate the natural phase assemblages in metabasics from Groix, a few broad conclusions can be drawn.

1. A glaucophane + garnet stability field could exist at higher temperatures than the breakdown of ferroglaucophane and lawsonite (see Hofmann, 1972, Nitsch, 1972)
2. The garnet forming reaction on the prograde metamorphic path could correspond with either reaction (1) or reaction (3) of Fig. 2.

If it is assumed that zone I metamorphic conditions on Groix fell within the glaucophane + lawsonite facies, then the up grade transition into zone II must have involved crossing this garnet forming reaction (see Fig. 3).

On a low P/T gradient, and only if reaction (1) is degenerate, it is possible to enter a garnet + lawsonite + glaucophane divariant field on the low temperature side of reaction (3). Since the garnet does not conform to the special compositional constraints enabling

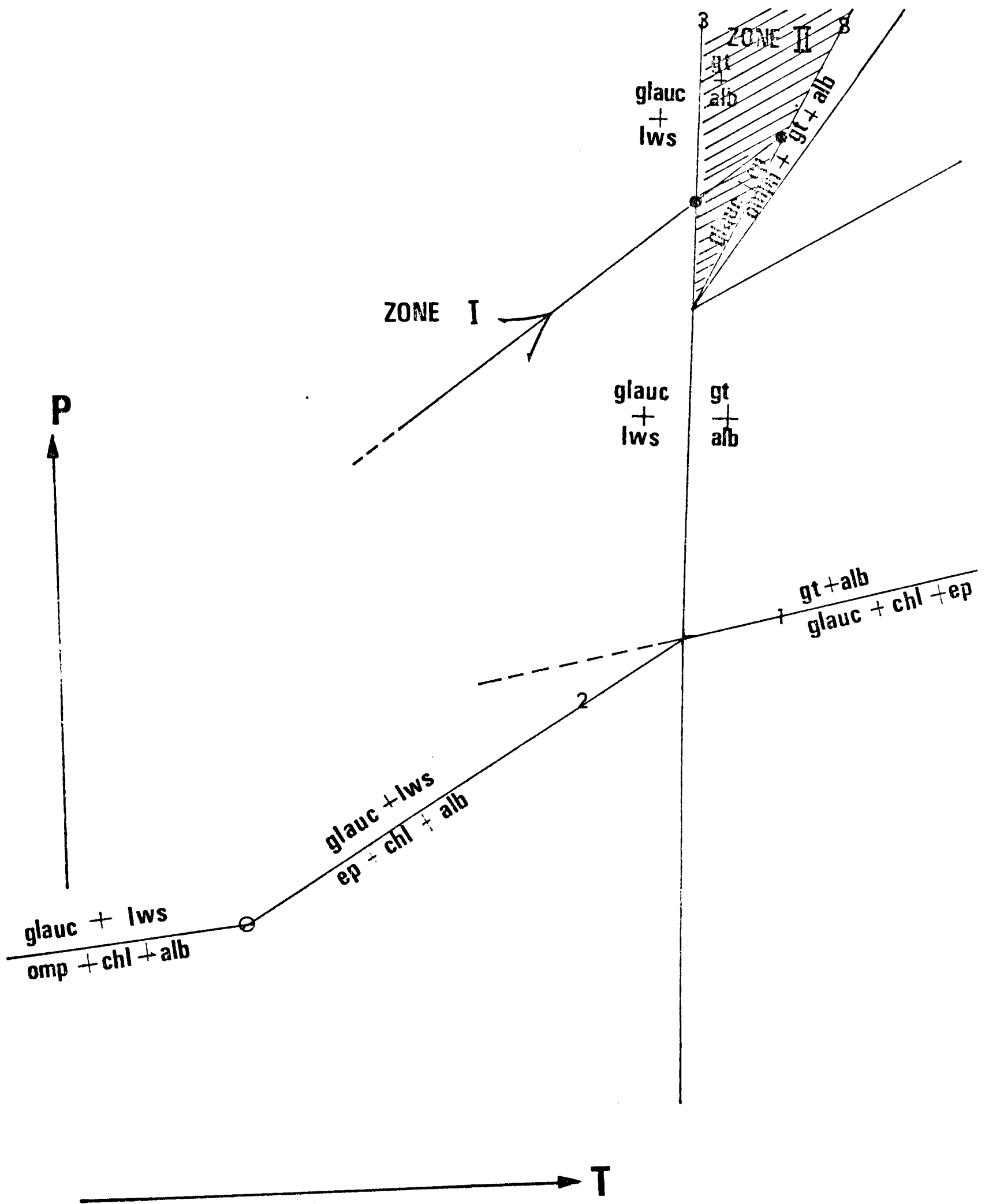


Fig. 4.3 Path of prograde metamorphism in Groix metabasites in relation to proposed petrogenetic grid.

degeneracy (no Ca; FM:A < chlorite), garnet and lawsonite cannot be stable together in a glaucophane producing field. This prediction is in agreement with the textural interpretation of lawsonite pseudomorphs as presented in Chapter 3 (section 6.7), whereby lawsonite is thought to have broken down before garnet growth.

With a rather higher P/T gradient, one can pass directly from a zone I assemblage into a glaucophane + garnet + epidote stability field by crossing reaction (3), if it is degenerate. In this case, the conditions for degeneracy are rather easier to meet for a wide range of bulk composition, since the grossular content of garnet falls near 20 mole percent (e.g. TUD5, PC10, LM12 garnet compositions in Appendix).

3. Continuing on the same P/T gradient path as proposed in Fig. 3, one traverses a glaucophane + garnet + epidote stability field (similar to that observed in the epidote zone of the Ouégoa schists, New Caledonia by Black, 1973) and eventually intersects equilibrium (8). Anywhere on this equilibrium, the univariant assemblage Glaucophane + epidote + omphacite + garnet + albite can co-exist, corresponding with the observed paragenesis of the present author's PC10. In reality, this equilibrium assemblage can occupy a divariant field in P,T space, due to solid solution of Fe + Mg in glaucophane, omphacite and garnet.

4. Retrogression from either zone I or zone II assemblages could occur by offloading at constant temperature. The upper pressure limits for chlorite + albite + epidote (greenschist) assemblages are set by reactions (2) and (1). Since reaction (3) is (chl)(ep), and has only a metastable extension into the chlorite + epidote producing field, there is no garnet-bearing greenschist subfacies.

In principle, actinolite is a stable, but non-essential phase of the greenschist facies here defined. Thoroughly retrogressed meta-

basics (see Section 7.2, Chapter 2) do contain minor amounts of actinolite, whereas partially retrogressed assemblages are typified by the occurrence of barroisite and magnetite.

4.7. Influence of grade and host rock chemistry on mineral chemistry

Shortcomings of the above approach lie in the fact that bulk rock chemistry (in particular variable Mg/Mg + Fe ratio) may control the composition and the appearance or disappearance of a phase at constant pressure and temperature, thus blurring the univariant reactions into divariant P,T bands.

In order to interpret the differing parageneses in terms of relative equilibrium conditions, it is very important to decide which changes in mineral chemistry are due to bulk chemical control and which are due to changing grade. With this end in view, there follows a summary of the variation of mineral composition in relation to host rock composition and metamorphic zone (as defined in Chapter 3). The chosen compositional parameter varies over a wide range (0.17 - 0.83) but all the rocks quoted can be described as metabasic schists (Calcareous, Basic and Magnesian schists of Mäkinen, 1967. Banded epidote schists and massive metabasic schists of the present author).

It can be readily appreciated from an inspection of Table 2.4, that, although there is a discernible and systematic influence of whole rock composition on phase composition, the variation in phase composition lies between certain limits for a given zone. This is particularly apparent for glaucophane, which shows a continuous solid solution across the zones, with the Mg content increasing with grade almost irrespective of co-existing ferromagnesian phases.

The incoming of garnet into metabasic assemblages is thus controlled by a 'sliding reaction', rather different from that envisaged as the model petrogenetic grid.

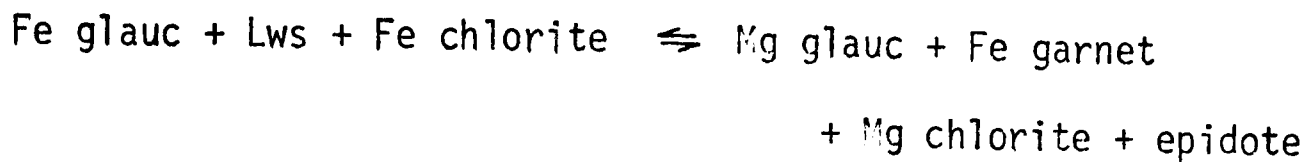
TABLE 2a

	Zone I		Zone II	
	Whole rock Mg/Mg + Fe < 0.5	Whole rock Mg/Mg + Fe > 0.5	Whole rock Mg/Mg + Fe < 0.5	Whole rock Mg/Mg + Fe > 0.5
Garnet	Alm ₅₈ Spess ₁₇ Gross ₁₀ Pyr ₅	Absent	Alm ₆₃ Spess ₄ Gross ₁₀ Pyr ₅	Alm ₅₄ Spess ₆ Gross ₂₆ Pyr ₇
Glaucophane	Feglauc ₅₀₋₆₀	Feglauc ₄₅₋₅₀	Feglauc ₃₀₋₄₀	Feglauc ₂₅₋₃₀
Epidote	Ps ₂₅₋₃₀	Ps ₂₁₋₂₅	Ps ₁₄₋₂₀	Ps ₂₀₋₂₅
Chlorite Fe/Fe + Mg	0.65	0.83 0.65	RETROGRESSIVE 0.54 0.65	
Actinolite	Absent	Trem ₆₃₋₇₁	Barroisite with 40-50% Trem	Barroisite with 45-54% Trem

Rocks used in the compilation of Table 2a

LH13		0.17
LH11		0.21
LH12		0.24
LM1		0.27
LM3		0.33
KM11	0.41	
PC10*	.	0.426
LM4		0.43
LH4		0.44
KB1*	0.44	
TUD5*		0.56
PN1*	0.62	
LM12*		0.63
LIN-1*	0.67	
BM1	0.75	
BIL3*	0.83	

*rocks described and analysed by present author. All others are due to Mekanjuola & Howie, 1972.



This exchange reaction involves all the important co-existing minerals of the metabasic rocks in both primary metamorphic zones on Groix, and as such may be regarded as a gross oversimplification - the actual reaction paths with increasing grade may differ from domain to domain within one thin section. It does, however, express a simple hypothesis - the reaction which brings garnet into the metabasic assemblages on Groix operated probably during, and definitely after, the breakdown of lawsonite (see section 3.6.7 of Chapter 3).

The reaction as quoted above is exactly similar to that given by Black (1973) to explain assemblage changes across a lawsonite zone/epidote zone isograd in the Ouégoa schists of New Caledonia. It is at least surprising, and encouraging, to find evidence for the occurrence of such an isograd on Groix.

If bulk chemistry had a very large influence on the position of the 'garnet in metabasics' isograd then one would not expect to observe a sharp transition in the field. Instead there should be a large area where garnet/epidote and glaucophane/lawsonite assemblages would be interbedded, garnet first appearing at the low temperature margin only in rocks with low $\frac{\text{Mg}}{\text{Mg}+\text{Fe}}$ ratios. By the same token, lawsonite would persist in rocks with high Mg/Mg + Fe and low $f\text{O}_2$ right up into the so-called 'epidote zone'.

On Groix, the 'garnet in metabasics' isograd as defined by the present author is a relatively sharp dividing line (see Fig. 1, Chapter 2) between the two prograde metamorphic zones, and the transition zone, if it is recognizable at all, is certainly less than 200 m wide (between Locmaria Plage and Kermarec on the south coast; between Les Viviers de Stauverec and Port Melin on the north coast). Such a narrow 'isograd zone' would suggest that for the normal range of basic rocks at constant

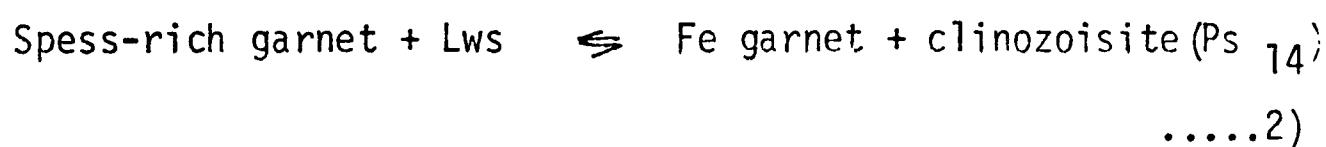
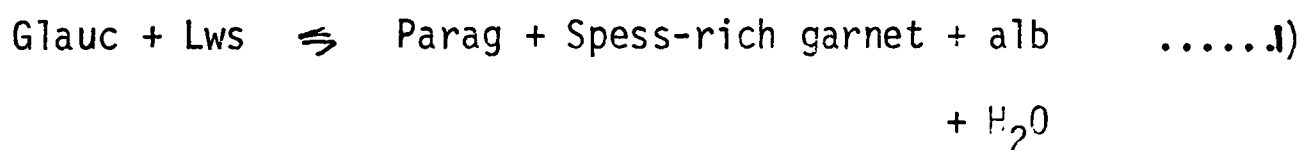
pressure, garnet becomes stable at a particular temperature and that this temperature may correspond with the breakdown of lawsonite.

As regards the variation of epidote composition in metabasic parageneses, the present author believes that the major controlling factor is changing fO_2 during prograde metamorphism (Liou, 1973) and this to some extent disguises the effect of changing bulk chemistry.

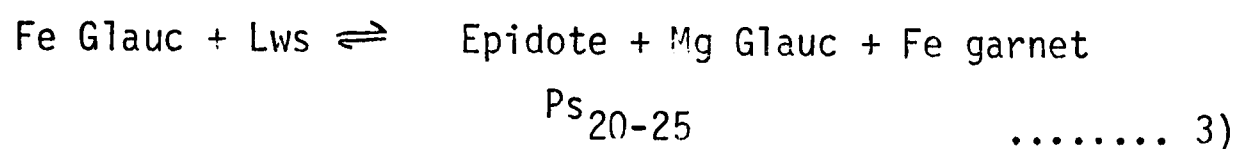
Since prograde metamorphism on Groix is a history of progressive dehydration with increasing temperature, one would expect fO_2 to fall with increasing grade as long as fluids were not buffered by the assemblages and evolved water of dehydration was free to escape.

Generally speaking, epidotes with high Fe^{3+} contents (Pistacite end-member ≈ 30 mole %) are restricted to rocks with high CaO contents (11.5 - 16 wt.%) in the low grade Zone I. In these rocks, Fe_2O_3/FeO is always high (> 1.0) and lawsonite has not been stabilized (see Felix, 1969).

By contrast, epidotes with low pistacite content ($Ps = 14$ mole %) are only found in those zone II assemblages with lawsonite pseudomorphs. As described in Chapter 3, section 6.5, the pseudomorphs consist of an aggregate of albite, paragonite and clinozoisite, with the clinozoisite often clustered towards the centre of the mass. These clinozoisites have been shown to have crystallized in equilibrium with small, inclusion free garnets whose growth post-dates the lawsonite breakdown (see analyses of PM1 garnet, paragonite and clinozoisite in Appendix III). The formation of the pseudomorphing assemblage, during prograde metamorphism, under the conditions prevailing in Zone II can be expressed by the following 'sliding' reactions



Reaction 1) would proceed at low temperatures, especially for rocks with low Mg/Mg+Fe, to produce a spessartine-rich garnet core. As the temperature increased, the garnet would become more Fe rich as lawsonite continued to break down. Eventually, Fe-rich garnet rims would be left in equilibrium with clinozoisite and all lawsonite would have disappeared. For rocks with higher Mg/Mg +Fe, a rather different sequence of 'sliding' reaction is envisaged, since the onset of lawsonite breakdown is at a higher temperature where spessartine-rich garnet is not stable.



In reaction 3) there is an exchange of Fe^{3+} between the low temperature glaucophane (say, 20 mole % riebeckite) and the epidote on the higher temperature side. Only if $f\text{O}_2$ falls dramatically with increasing grade can clinozoisite (Ps14) be formed in such a reaction, while O_2 is released in the R.H.S.

Thus we can conclude that Mg/Mg+Fe of the rock has little or no effect on the composition of the epidote produced, beyond controlling the temperature at which lawsonite breakdown can occur. The higher the temperature, the more likely are low $f\text{O}_2$ conditions due to ongoing dehydration reactions.

4.8. Estimation of pressure, temperature and fluid phase composition during metamorphism of Groix metabasics

The glaucophane eclogites (e.g. the present author's specimens PC10, LM12. Mäkanjuola's LM1, PYM1, LH3, LH6, LH16 and Triboulet's CdS11, LM7) have been selected as those rocks most likely to record the equilibrium conditions attained during the peak of Zone II metamorphism on Groix. This conclusion has been arrived at after due consideration of the structural and textural evidence relating to mineral

growth episodes (see Chapter 2, Table 2) and the predictions derived from model petrogenetic grids (see Fig. 3, this chapter).

Pressure/temperature conditions have been inferred from the activity of jadeite in omphacite co-existing with quartz and albite and from the $\text{Fe}^{\text{II}}/\text{Mg}$ partitioning between adjacent garnet and omphacite rims (Råheim and Green, 1975). Limits have been placed on the fluid phase composition using the known X_{CO_2} -Temp stability of lawsonite at 7 kb (Nitsch, 1972) and the P_{CO_2} -Temp equilibrium curve for the Rutile + Quartz + Calcite $\rightleftharpoons$ Sphene + CO_2 transition (Schuiling and Wink, 1967).

Further checks on the inferred $P, T, X_{\text{H}_2\text{O}}$ conditions are provided by the Si^{4+} content of phengites (Velde, 1965) and the occurrence of ferroglaucophane in Zone I (Hoffmann 1972).

Finally, equilibrium metamorphic temperatures derived from oxygen isotope studies in the Franciscan (Taylor and Coleman, 1968) and New Caledonian (Black, 1974) blueschist belts have proved to be of great value in assessing the validity of Fe/Mg partition geothermometers at low temperatures.

4.9. Application of Råheim and Green's (1974) Fe/Mg partition geothermometer

Råheim and Green (1974) have presented experimental evidence to show the dependence of $K_D^{\text{ga-cpx}}$ (the Fe/Mg partition coefficient between garnet and clinopyroxene) on pressure and temperature for a range of basaltic compositions ($\text{Mg}/\text{Mg}+\text{Fe} \times 100 = 6.2 - 93$). They conclude that, despite discrepancies with theoretically expected K_D values based on available thermodynamic data (Banno, 1970), there is a good linear relationship of $\ln K_D^{\text{ga-cpx}}$ with $\frac{1}{T}$ at 30 kb. At this pressure, $K_D^{\text{ga-cpx}}$ values ranged from 18.8 ± 2 at 600°C to 1.45 ± 0.1 at 1400°C , and were arrived at by microprobe analysis of phases grown

in basaltic glasses at very low fO_2 in iron capsules.

The pressure dependence of the K_D^{ga-cpx} value was determined in the range 20-40 kb at constant temperature (1100°C) and found to be directly proportional to $\ln K_D$, though the pressure coefficient is greater than theoretically predicted from volume data (Banno, 1970).

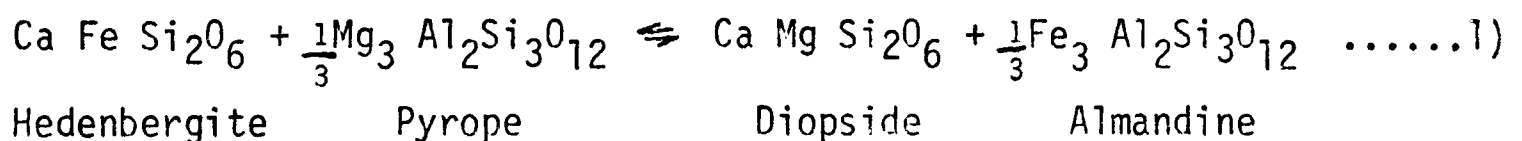
In addition, changes (Mg/Mg+Fe x 100) in the bulk composition from 6.2 - 76.6 were found to have no detectable influence on the K_D^{ga-cpx} measured at 1000°C, 30 kb.

Combining the experimental data defining the pressure, temperature and compositional effects on K_D^{ga-cpx} , Raheim and Green derive an empirical relationship

$$T (^{\circ}K) = \frac{a + \frac{P^*}{R} (P - P_0)}{\ln K_D^{ga-cpx} - b} = \frac{3686 + 28.35 \times P \text{ (kb)}}{\ln K_D + 2.33}$$

where a and b are constants which can be determined from the experimental data and p^* is the pressure coefficient theoretically corresponding to ΔV (molar volume change of the exchange reaction) but in practice derived from the slope $(\frac{\Delta V}{RT})$ of the linear $\ln K_D$ vs Pressure relationship at constant T.

Considering the relevant exchange reaction, using $\frac{1}{3}$ formula units for garnet,



then at equilibrium, the equilibrium constant is defined as

$$K = \frac{(a_{\text{Diop}}^{\text{cpx}})(a_{\text{Alm}}^{\text{ga}})}{(a_{\text{Hd}}^{\text{cpx}})(a_{\text{Pyr}}^{\text{ga}})} \quad \text{.....2)}$$

where $a_{\text{diop}}^{\text{cpx}}$ is the activity of diopside component in the clinopyroxene solid solution and $a_{\text{Alm}}^{\text{ga}}$ is the activity of almandine in the garnet. Similarly, $a_{\text{Hd}}^{\text{cpx}}$ and $a_{\text{Pyr}}^{\text{ga}}$ refer to the activities of hedenbergite and pyrope in their respective solid solutions.

Now, the activity of the i^{th} component in a particular phase is related to the chemical potential μ_i^0 at the standard state compared with the potential μ_i at the temperature of equilibrium, T .

$$\mu_i = \mu_i^0 + RT \ln a_i \quad \dots\dots\dots 3)$$

If we assume ideal ionic mixing on one site for the end members in the equilibrium solid solution, then

$$a_i = X_i \quad \dots\dots\dots 4)$$

where X_i is the mole fraction of component i derived from chemical analysis of the phase, and recalculation of the formula unit, assuming perfect stoichiometry.

Thus, in the specific case of reaction 1) K will become

$$K = \frac{X_{\text{Fe}}^{\text{gt}} / X_{\text{Mg}}^{\text{gt}}}{X_{\text{Fe}}^{\text{cpx}} / X_{\text{Mg}}^{\text{cpx}}} = K_D^{\text{ga-cpx}} \quad \dots\dots\dots 5)$$

Since, although garnet has three sites, only $\frac{1}{3}$ formula units are used.

Theoretically, the dependence of $\ln K$ on pressure and temperature are given as follows:-

$$\left(\frac{\ln K}{T} \right)_P = \frac{\Delta H}{RT^2} \quad \dots\dots\dots 6)$$

$$\left(\frac{\ln K}{P} \right)_T = \frac{\Delta V}{RT} \quad \dots\dots\dots 7)$$

Unfortunately, enthalpy of formation or solution (ΔH) is not known for hedenbergite or almandine so it is not possible to compare the empirical relationship of $\ln K_D$ against $1/T$ given by Raheim and Green (1974) with theoretical expectations.

With equation (7), it is possible to compare directly the molar volume change ΔV of the reaction (1) with the p^* , or pressure coefficient of the $\ln K_D$ vs. P experimental relationship. It appears that p^* is

very much greater (2.357cc) than the value calculated from end-member molar volume data ($\Delta V = 1.55 \text{ cc} - 0.91 \text{ cc}$, depending on choice of formula volumes).

This discrepancy is probably due to excess volumes of mixing in one or other or both of the solid solutions under consideration. In addition, there is the problem of grossular component in the garnets, which may enhance the molar volume changes in the 'natural' system.

The occurrence of grossular component in natural garnet solid solutions presents the greatest single obstacle to an application of R  heim and Green's geothermometer to natural eclogites. This is because the mixing of grossular and pyrope in the garnet solid solution has been found to be distinctly non-ideal, with a positive deviation increasing with decreased temperature (Hensen et al., 1975).

The effect of this non-ideality is to lower drastically the K value at equilibrium since $a_{\text{pyr}} = x_{\text{Mg}}^{\text{gt}} \cdot \gamma_{\text{pyr}}$ where γ_{pyr} is an activity coefficient largely controlled by the Grossular-Pyrope and Almandine-pyrope interactions (Ganguly and Kennedy, 1974).

$$K = K_D^{\text{ga-cpx}} \cdot \frac{(\gamma_{\text{Alm}}/\gamma_{\text{Pyr}})}{(\gamma_{\text{Hd}}/\gamma_{\text{Diop}})} \dots\dots\dots 8)$$

Thus, the R  heim and Green geothermometer is strictly valid only if $\frac{\gamma_{\text{Alm}}/\gamma_{\text{Pyr}}}{\gamma_{\text{Hd}}/\gamma_{\text{Diop}}}$ is constant with varying P,T and phase composition. This could only occur if $x_{\text{Gross}}^{\text{Gt}}$ were kept constant and $\gamma_{\text{Alm}}/\gamma_{\text{Pyr}}$ did not change with varying Mg/Mg+Fe bulk composition. Such a combination of circumstance is unlikely in view of the fact that activity coefficients are composition and temperature dependent. It is just possible that co-variance of $\gamma_{\text{Alm}}/\gamma_{\text{Pyr}}$ and $\gamma_{\text{Hd}}/\gamma_{\text{Diop}}$ may cancel out if the interchange energies of the Almandine-pyrope and hedenbergite-diopside joins are of similar magnitude and both positive (Oka and Hatsumoto, 1974).

If this were so, then Grossular-pyrope non-ideal mixing would bring about the largest effect on K.

$$\ln K_D^{\text{ga-cpx}} = \ln K + \frac{W_{12}}{RT} (X_{\text{Fe}} - X_{\text{Mg}})^{\text{Gt}} - \frac{W_{\text{HdDi}}}{RT} (X_{\text{Hd}} - X_{\text{Di}})^{\text{cpx}} \\ - \frac{(W_{13} - W_{23})}{RT} (X_{\text{Ca}})^{\text{Gt}} \dots\dots\dots 9)$$

where W_{12} is the almandine-pyrope interchange energy

W_{HdDi} is the hedenbergite-diopside interchange energy

W_{13} is the almandine-Grossular interchange energy

W_{23} is the pyrope-grossular interchange energy

Interchange energies are related to activity coefficients in the following way:

$$Rt \ln \gamma_i = W_{ij} (1 - X_i)^2$$

This formulation of activity coefficients is for a random distribution of components i and j in a binary solution with mixing on one site (Guggenheim, 1952). The departure from ideal, ($\gamma_i = X_i$), mixing is said to be regular, and implies a regular variation of ΔH_{excess} and ΔS_{excess} with composition at constant temperature.

The implications of this departure from ideality for grossular-pyrope mixing are summarised in Fig. 4, where data from Raheim and Green's 600°C and 700°C runs at 30 kb are used, in conjunction with the W_{23} value proposed by Hensen to calculate the theoretical $\ln K$ values and compare this with the observed $\ln K_D^{\text{ga-cpx}}$.

Also, data from the Puerto Cabello eclogite 1059 (Morgan, 1970) and the present author's glaucophanic eclogite PC10 (OMD - GTA omphacite/garnet pair) are used, assuming temperatures of equilibration of 525°C and 400°C respectively.

W_{23} is assumed to be temperature dependent according to the equation given in Hensen et al. (1975).

$$W_{23} = 7460 - 4.3 T (^{\circ}\text{K}) \text{ Cal/mole}$$

whilst, all the other W values are assumed to be constant over the temperature range $400^{\circ} - 700^{\circ}\text{C}$.

$$W_{12} = 2980 \pm 500 \text{ cal/mole (Ganguly & Kennedy, 1974)}$$

$$W_{13} = 1350 \quad (\text{Ganguly & Kennedy, 1974})$$

$$W_{\text{HdDi}} = 2390 \quad (\text{Oka & Matsumoto, 1974})$$

Then if all the measured mole fractions listed below are substituted into equation (9) one can calculate the theoretical $\ln K$ values given

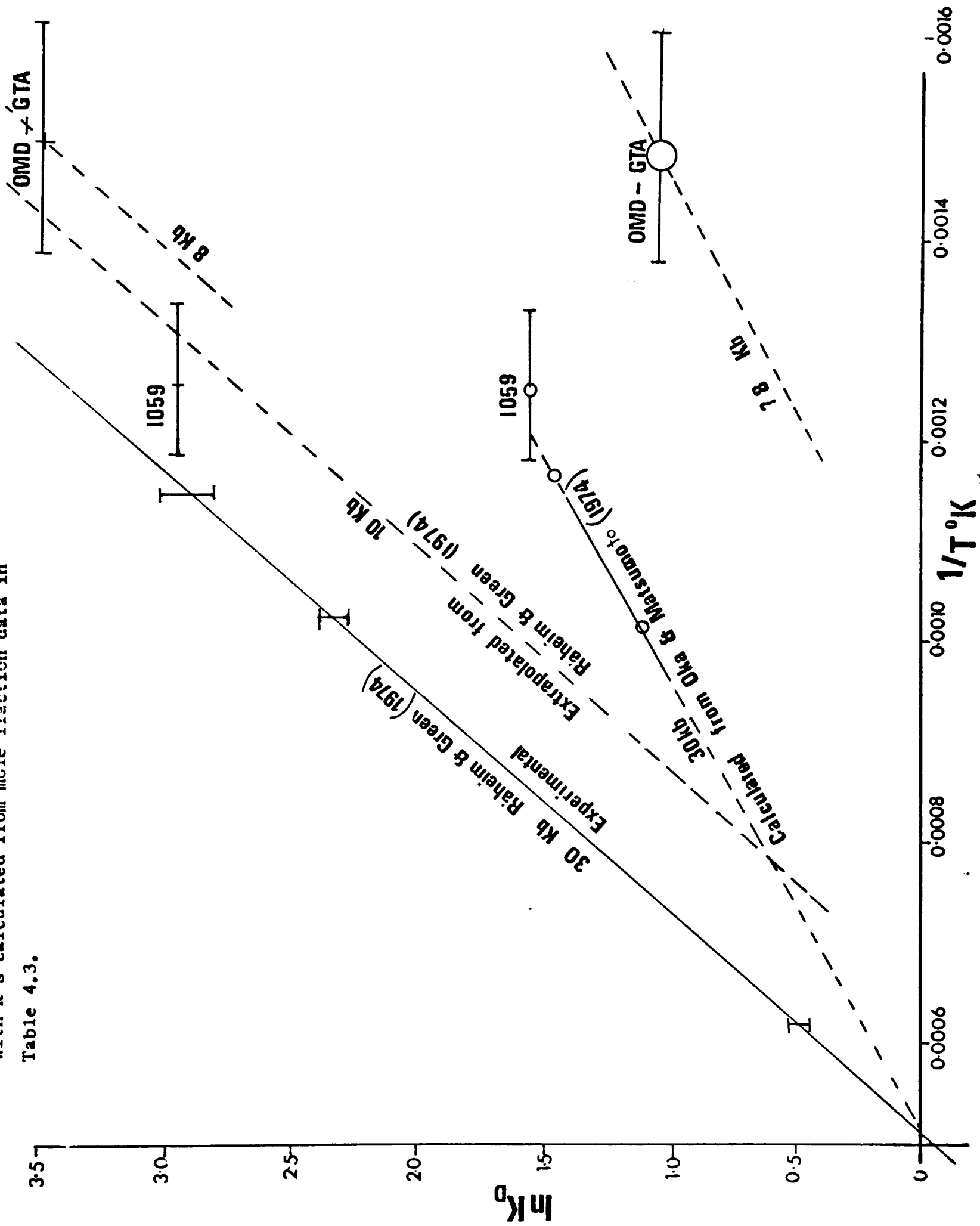
TABLE 3

	400°C	525°C	600°C	700°C
$x_{\text{Fe}}^{\text{Gt}}$	0.623	0.476	0.62	0.60
$x_{\text{Mg}}^{\text{Gt}}$	0.0754	0.2033	0.03	0.10
$x_{\text{Ca}}^{\text{Gt}}$	0.251	0.232	0.355	0.30
$x_{\text{Hd}}^{\text{Cpx}}$	0.112	0.05	0.31	0.25
$x_{\text{Di}}^{\text{Cpx}}$	0.429	0.42	0.27	0.36
$\ln K_D^{\text{ga-cpx}}$	3.454	2.978	2.934	2.35
$\ln K$	1.072	1.58	1.48	1.14

in the last line of Table 3 and compare them with the experimentally determined $\ln K_D^{\text{ga-cpx}}$ which assume ideal mixing.

It can be seen from Fig. 4 that the main effect of an increasing grossular-pyrope non-ideality with decreasing temperature, as long as $x_{\text{Ca}}^{\text{Gt}}$ is kept within the limits 0.25 - 0.35, is a decrease in the graph slope ($-\frac{\Delta H}{RT^2}$). At high temperatures, (ca. 1600°C), as $K_D^{\text{ga-cpx}}$ approaches 1.0 at 30 kb, the grossular-pyrope non-ideality of mixing

Fig. 4.4 Comparison of K_d 's observed at constant pressure, with K 's calculated from mole fraction data in Table 4.3.



disappears and the extrapolated $\ln K$ (theoretical, modelled for $x_{\text{Gt}}^{\text{Ca}} = 0.3$) vs $1/T$ relationship coincides with Raheim and Green's empirical relationship extrapolated to high temperatures.

So, in conclusion, Raheim and Green's empirical relationship fits the experimental data for Fe/Mg partitioning between garnet and pyroxene quite well in the range $600^{\circ}\text{C} - 1400^{\circ}\text{C}$ and 20 - 40 kb for glasses of basaltic composition but highly variable $(\text{Mg}/\text{Mg}+\text{Fe}) \times 100$. Whether this experimentally determined relationship can be used to derive the equilibrium temperatures of natural eclogites depends on:-

1. A knowledge of the mixing properties of all the important end-member components present in natural garnet and pyroxene solid solutions. As a zeroth approximation, the grossular-pyroxene solid solution can be taken as the major source of non-ideality here, and the effect of this on $\ln K$ can be modelled accordingly.
2. A reasonable estimate of pressure during equilibration. This can be achieved by measuring the jadeite contents of pyroxene in equilibrium with quartz and albite in the assemblage of interest. If plagioclase is not present in the assemblage (as with the Puerto Cabello, Californian and New Caledonian eclogites) then only a minimum pressure can be established, and the temperature is not constrained.
3. Accuracy of analysis of equilibrium mineral pairs. The wet chemical analyses generally available (see Coleman et al., 1965) may not be representative of any one equilibrium composition, but instead may be gross averages for heterogeneous grains. In that case, only rim compositions derived from the microprobe analysis of adjacent grains can be used to characterise the temperature at the end of mineral growth, as long as retrograde effects are avoided.

4. Whole rock composition of the eclogite concerned. Although Raheim and Green have demonstrated that variable $(\text{Mg}/\text{Mg}+\text{Fe}) \times 100$ has no effect on $K_D^{\text{ga-cpx}}$ at constant pressure and temperature, the effect of other minor components such as Na_2O and K_2O is not known. As long as the eclogite concerned has a bulk chemistry near that of the quartz tholeiite 'B' of Green and Ringwood (1967), then the experimentally determined K_D 's can be directly compared with naturally observed K_D 's.

TABLE 4

	Tholeiitic glass (37.4)	Glaucophane eclogite PC10 (42.7)	Tholeiitic glass (69.4)	Glaucophane eclogite LM12 (63)
SiO_2	51.1	47.02	54.2	49.38
TiO_2	1.8	2.26	1.8	1.48
Al_2O_3	14.3	13.99	15.0	14.36
FeO_T	14.7	15.908	6.6	11.69
MnO	0.15	0.23	0.15	0.2
MgO	4.9	4.95	8.4	7.44
CaO	9.8	10.63	10.0	9.64
Na_2O	2.6	3.52	3.0	4.48
K_2O	0.7	0.34	1.0	0.03
P_2O_5	0.2	0.33	0.2	0.10
	100.2	99.18	100.3	98.81

The numbers in brackets refer to the $(\text{Mg}/\text{Mg}+\text{Fe}) \times 100$. FeO_T is total iron assigned as FeO .

Table 4 compares the compositions of the glaucophanic eclogites analysed by the present author with two appropriate starting materials used by R  heim and Green.

4.10. Extrapolation of R  heim and Green's geothermometer to temperatures less than 600  C at pressures less than 20 kb

Since R  heim and Green's geothermometer depends on an empirical relationship extracted from experimental data, there is no a priori reason to support its validity outside the range of 600  C - 1400  C and 20 - 40 kb.

With decreasing temperature, even with constant composition and pressure, other sources of non-ideality such as almandine-pyrope and diopside-jadeite interactions may become non-negligible and thus effect the $\ln K_D$ in a non-linear way.

Unfortunately, nothing is known of the temperature dependence of these interaction parameters (W_{12} , W_{13} , $W_{Hd.Di}$, $W_{Jd.Di}$) so it is not possible to model their influence on $\ln K_D^{ga-cpx}$ beneath 600  C.

The pressure dependence of $\ln K_D^{ga-cpx}$ is unlikely to change for lower pressures (say 10 kb) due to the incompressibility of solid phases. Thus, although the experimentally determined p^* is greater than the ideal ΔV reaction, it is probably justifiable to assume $p^* - \Delta V = \text{constant}$ over a large pressure range.

If this last assumption is correct, then it is possible to predict the $\ln K_D^{ga-cpx}$ values at 10 kb for experimental runs at 600  C and 700  C by applying the equation:

$$\ln K_{30 \text{ kb}} = \frac{p^*}{RT} (P - P_0) + \ln K_{10 \text{ kb}}$$

where $P - P_0 = 20 \text{ kb}$, $p^* = 2.357 \text{ cc}$

$R = 83.147 \text{ cc.bar/deg mole}$.

This extrapolation to 10 kb is illustrated in Fig. 5 as a solid line

between 600°C and 700°C, and would be applicable to natural eclogite assemblages with garnets in the compositional range 0.25 - 0.35 mole fraction grossular and pyroxenes with $(^{2\text{Na}}/2\text{Na} + \text{Fe} + \text{Mg} + \text{Ca}) \times 100$ near 42.

The dashed line in Fig. 5 marks the linear extension of the 10 kb equilibrium partition relationship down to 400°C.

For the purpose of comparison only, various type B and C eclogites (Coleman et al., 1965), for which independent estimates of pressure and temperature are available, are plotted on Fig. 5. The data are listed in Table 5.

It can be seen immediately that there is a surprisingly good agreement within error between temperatures estimated from a variety of different methods, and temperatures calculated from the Råheim and Green relationship at 10 kb. This agreement is surprising in view of the fact that errors in pressure and temperature estimation are additive, and that grossular content of the garnets varies from 12.3 mole % in the case of the Ward Creek type IV exotic block to 25.8% in the case of the New Caledonian exotic block.

For the Ile de Groix glaucophanic eclogites, with pressure estimates of 8 kb (see next section and Table 6), the Råheim and Green geothermometer gives $400 \pm 50^\circ\text{C}$ as a possible temperature of equilibration. Such a temperature is comparable with oxygen isotope temperatures quoted for the lawsonite-epidote transitional zone in New Caledonia (Black, 1974, Bottinga and Javoy, 1975: specimen 31 NC 62 Quartz-ilmenite pair $T_e = 417^\circ$, Garnet-ilmenite $T_e = 336^\circ\text{C}$). Furthermore, the grossular content of the garnets from Groix glaucophanic eclogites is fairly uniform around 25 mole %, and the $(2\text{Na}/2\text{Na} + \text{Fe} + \text{Mg} + \text{Ca}) \times 100$ ratio in the co-existing omphacites is very similar to that in pyroxenes from Råheim and Green's 700°C run at 30 kb (i.e. 42).

If we assume a 10% uncertainty in the $\ln K_D^{\text{ga-cpx}}$ vs $1/T$ relation-

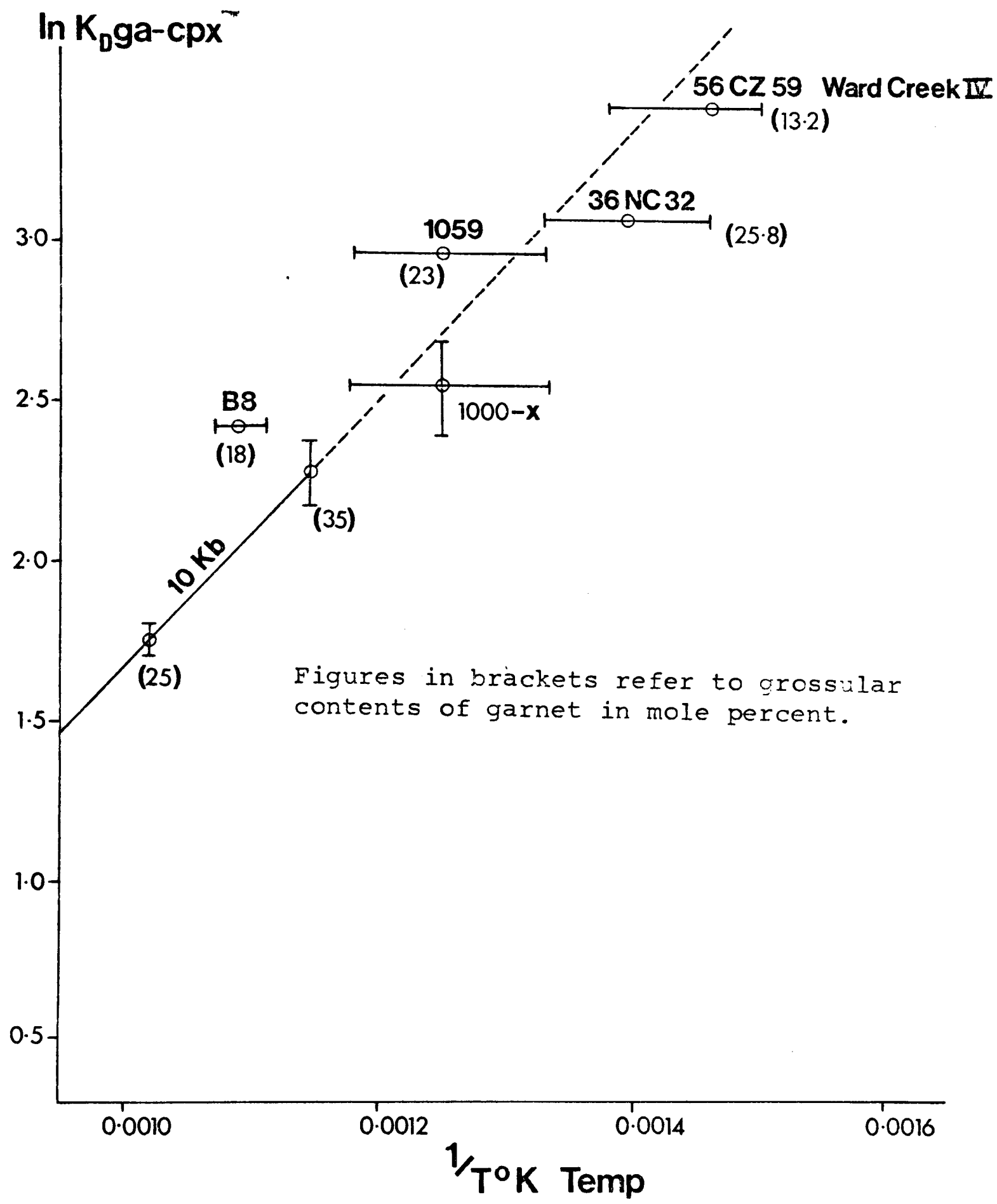


Fig. 4.5 Extrapolation of Raheim and Green's geothermometer to temperatures less than 600°C.

TABLE 5

	K_D ga-cpx	P Kbar estimate	$T^{\circ}\text{C}$ estimate	$T^{\circ}_{\text{calc C}}$ Raheim & Green
62 RGC 58 ⁽¹⁾ Ward Creek type IV	30.99	10	$410 \pm 50^{(2)}$	410
36-NC-62 ⁽³⁾ New Caledonia	21.67	10	$442 \pm 35^{(4)}$	462
1059 ⁽⁵⁾ Puerto Cabello Venezuela	19.66	10 min	$525 \pm 50^{(6)}$	475
1000-X ⁽⁷⁾ Puerto Cabello Venezuela	13 ± 2	10 min	586 (8)	550
B8 ⁽⁹⁾ Bourron, France	11.4	10 min	$640 \pm 15^{(10)}$	580

- (1) Coleman et al. 1965.
- (2) Quartz-Glaucophane $\delta^{18}\text{O}$ temperature from 56-CZ59. Taylor & Coleman, 1968.
- (3) Coleman et al. 1965.
- (4) Average of Quartz-Muscovite and Quartz-Glaucophane $\delta^{18}\text{O}$ temperatures. Taylor & Coleman, 1968.
- (5) Morgan, 1970.
- (6) Calcite-dolomite temperatures, Morgan, 1970.
- (7) Morgan, 1970. Quoted in Vogel & Garlick, 1970.
- (8) Quartz-clinopyroxene $\delta^{18}\text{O}$ temperature in Bottinga & Javoy, 1975.
- (9) Velde, 1970.
- (10) Quartz-garnet and quartz-clinopyroxene $\delta^{18}\text{O}$ temperatures in Bottinga and Javoy, 1975.

ship as determined by R  heim and Green (1974), then squeezing the 30 kb line through the extremes of the $\ln K_D^{ga-cpx}$ error brackets at 600  Cm we obtain $T_{e \max} = 615^\circ\text{C}$, $T_{e \min} = 578^\circ\text{C}$. This $\pm 15^\circ\text{C}$ error on T_e is due mostly to uncertainties in the measurement of K_D in run 4549 (600  C, 30 kb). Errors on this run could combine with the errors in determining K_D 's for the Groix garnet-omphacite pairs (i.e. Fe^{3+} assignment, compositional zoning) to produce a $\pm 25^\circ\text{C}$ uncertainty in the temperature estimate, assuming a linear extrapolation of the $\ln K_D$ vs $1/T$ relationship to low T and P. To this may be added the probable errors in thermocouple measurements ($\pm 10^\circ\text{C}$), making a maximum precision of $\pm 35^\circ\text{C}$ for the R  heim and Green geothermometer.

An estimate of accuracy is far more difficult, since there is no error-free way of calibrating the low T end of the geothermometer. However, it must be stressed that, even if the actual temperature estimates turn out to be incorrect, statements concerning relative temperatures will continue to be valid.

4.11. Estimation of load pressure ($= P_{\text{TOTAL}}$) during metamorphism of the Groix glaucophanic eclogites

Essene and Fyfe (1967) were among the first to recognise that omphacite, as defined by Deer, Howie and Zussman (1963) is commonly developed in greenstones and metasediments of the glaucophane schist facies. This view is strongly supported by the occurrence of omphacite veins cutting Franciscan 'in situ' glaucophane-lawsonite schists (e.g. Type III metabasalts, Ward Creek, California. Coleman and Lee, 1963).

Furthermore, the low temperature/high pressure conditions of formation of these schists is attested by the oxygen isotope studies of Taylor and Coleman (1968), the occurrence of aragonite rather than

calcite and the highly ordered P2 symmetry of the omphacites (Clark and Papike, 1966).

Although there is no aragonite present on Groix, and the space group symmetry of the pyroxenes has not been determined, the present author feels it is justifiable to consider the development of omphacite in the metabasic schists as a prograde phenomenon taking place in situ at temperatures just in excess of the lawsonite breakdown (Nitsch, 1972). The reasons for this interpretation are outlined in Chapter 3 section 6.7 and further supported by the Fe/Mg partitioning data described above.

In order to assess the equilibria formation pressure of an omphacite pyroxene, it is necessary to have

- a) an accurate determination of the Jadeite + Quartz $\rightleftharpoons$ Albite equilibrium in the temperature range of interest.
- b) both a plagioclase of known composition and quartz present in the equilibrium assemblage.
- c) an estimate of equilibrium temperature (T).
- d) an estimate of the activity of jadeite component (a_{Jd}^{cpx}) in the pyroxene, derived from the recalculation of probe or wet chemical data and the application of some multi-component mixing model.

If all these requirements are met, and if we assume negligible compressibility, or thermal expansion of the solid phases, then at equilibrium:-

$$RT \ln a_{Jd}^{cpx} / a_{Alb}^{plag} = \Delta V \cdot \Delta P \quad \dots\dots\dots 1)$$

where R is the gas constant = 83.147 cc. bar/deg mole

a_{Alb}^{plag} is the activity of albite component in the plagioclase; in this study pure albite composition is assumed so $a_{Alb}^{plag} = 1.0$

ΔV is the molar volume change of the pure end-member reaction Jadeite + Quartz $\rightleftharpoons$ Low Albite; in this study $\Delta V = 16.98$ cc (data from Robie & Waldbaum, 1968).

ΔP is the pressure increment in bars from the equilibrium pressure of the end-member reaction at T.

4.12. Jadeite + Quartz $\rightleftharpoons$ Albite equilibrium

A hydrostatic determination of this equilibrium by Hays and Bell (1972) gives 16.4 ± 0.5 kb as the equilibrium pressure at 600 ± 5 °C. This is well within error of the results obtained by Newton and Smith (1967) and Boettcher and Wyllie (1968b) based on piston-cylinder apparatus, and agrees with an interlaboratory comparison of piston-cylinder calibration (Johannes et al., 1971) which gives an average preferred value of 16.3 ± 0.5 kb at 600°C.

For the purposes of this study, the albite breakdown was fixed at 16.3 kb at 600°C, and the (dP/dT) was assumed to be 20 bars/°C (see Fig. 6). The (dP/dT) at 400°C was calculated from the 3rd law entropies at 400°C and the molar volumes at S.T.P. of low albite, quartz and jadeite, using the data in Robie & Waldbaum (1968) and the Clausius-Clapeyron equation

$$\left(\frac{dP}{dT}\right)_{400^{\circ}\text{C}} = \frac{\Delta S}{\Delta V} = 20.5 \text{ bars/}^{\circ}\text{C}$$

Applying a correction for the effect of thermal expansion and compressibility, Popp and Gilbert (1972) calculate a slope of 20.2 bars/°C at 400°C and 12.3 kbar for this reaction, which is well within the uncertainties of the experimentally determined curve of albite breakdown.

At the very most, uncertainties in the assumed $\left(\frac{dP}{dT}\right)_{400^{\circ}\text{C}}$ could contribute ± 0.5 Kb error to the pressure estimate.

Fig. 4.6 Pressure/temperature conditions of equilibration for garnet/omphacite pairs (this study) from Groix, in comparison with other eclogites and pyroxene-bearing metabasalts.

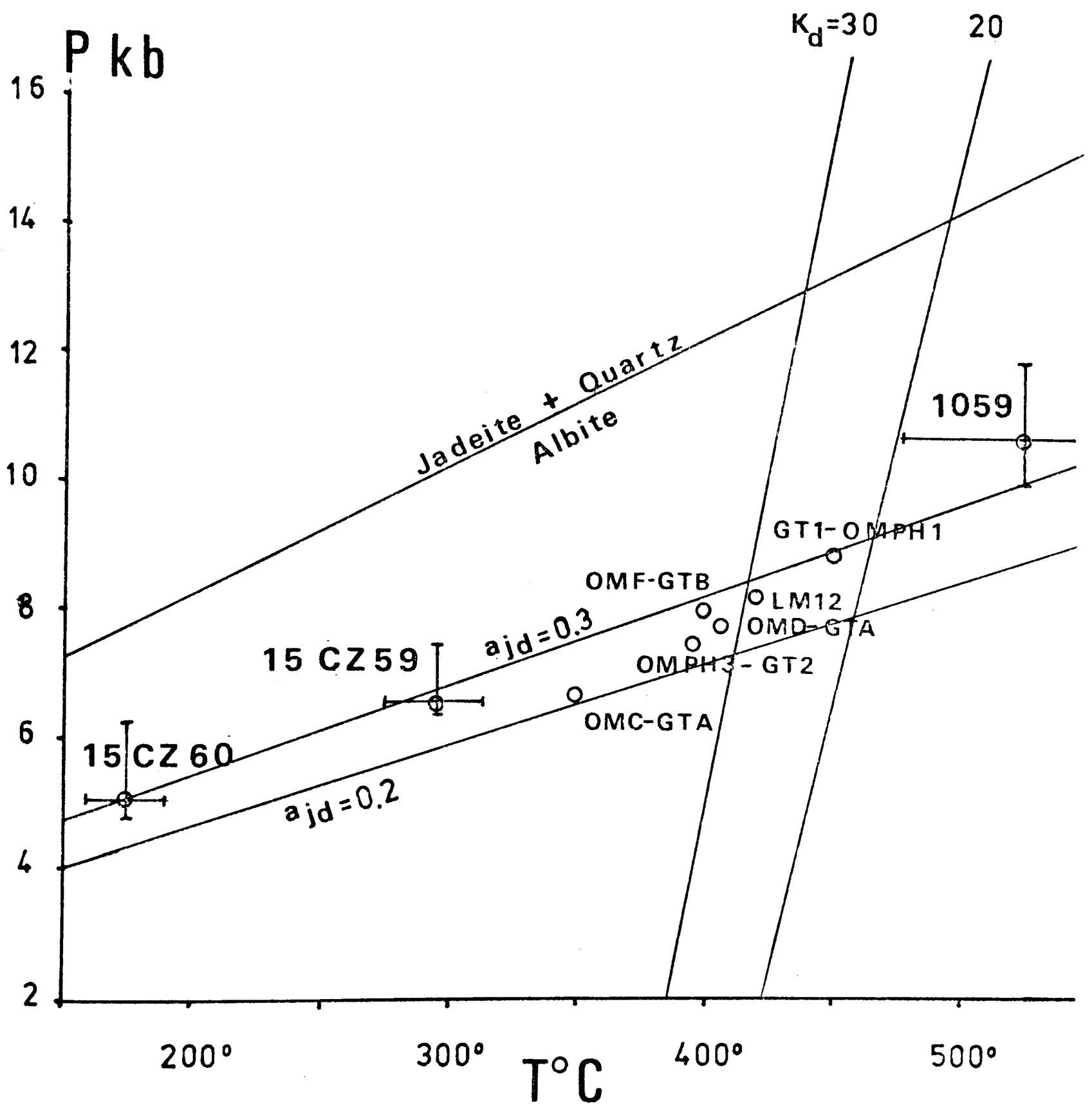


TABLE 6

Data used in geothermometry and geobarometry of glaucophanic eclogites

	X _{Di}	X _{Hd}	X _{CaTs}	X _{Ac}	X _{Jd}	T _e ^{oC}	Pe Kbar	A _{Jd}	K _D ^{ga-cpx}	Reference and comments
PC10										This study
OMC-GTA	0.4455	0.059	-	0.223	0.2955	350	6.7	0.22	62.5	)
OMPH3-GT2	0.4231	0.0936	-	0.1835	0.2832	390	7.5	0.217	35.3	)
OMD-GTA	0.429	0.1117	-	0.168	0.3139	410	7.8	0.243	31.6	)
OMF-GTB	0.4105	0.089	-	0.175	0.3439	400	8.0	0.273	34.1	)
OMPH1-GT1	0.3021	0.1091	-	0.2149	0.343	450	8.9	0.291	22.0	) Adjacent quartz only
LM12										
GT1-OMPH2	0.42	0.14	-	0.079	0.345	420	8.3	0.270	25.0	No quartz or albite
LM1	0.328	0.087	0.119	0.155	0.285	410	7.5	0.23	31.3	Makanjuola & Howie, 1972. Quartz bearing "
PYM1	0.394	0.112	0.036	0.157	0.288	450	8.0	0.223	21.6	"
LH3	0.350	0.088	0.043	0.193	0.313	430	8.0	0.253	25.6	No Quartz
LH6	0.328	0.089	0.08	0.152	0.32	410	8.0	0.26	29.9	Quartz bearing
LH9	0.418	0.085	0.07	0.145	0.372			0.297	-	73% modal omphacite
LH16	0.321	0.066	0.072	0.244	0.268	400	7.25	0.219	28.5	Quartz bearing
LH12	0.385	0.096	0.06	0.142	0.365			0.294	-	" "
CdS11	0.418	0.098	0.066	0.154	0.267	410	7.2	0.206	32.4	Triboulet 1974
LM7	0.261	0.146	0.067	0.226	0.292	450	8.3	0.245	19.5	No quartz or albite

continued

Table 6 (continued)

	X _{Di}	X _{Hd}	X _{CaTs}	X _{Ac}	X _{Jd}	T _e ^{oC}	P _e Kbar	A _{Jd}	K _D ^{ga-cpx}	Reference and comments
62 RGC 58 Ward Creek	0.308	0.123	0.087	0.205	0.277	410	7.6	0.241	31.0	Coleman et al. 1965
36-NC-62 New Caledonia	0.401	0.10	0.048	0.10	0.351	460	9.3	0.29	21.7	Coleman et al. 1965
1059 Puerto Cabello Venezuela	0.42	0.05	0.02	0.02	0.49	475	10.7	0.42	19.7	Morgan, 1970

Microprobe analyses from this study are recalculated by the Powell & Powell (1974) method. All other analyses are by wet chemistry, recalculated on the basis $Al^{iv} = Na - Fe^{3+}$. T_e is the equilibrium temperature derived from the Raheim & Green geothermometer (1974) and P_e is the minimum equilibrium pressure assuming quartz in excess, using the Ganguly mixing model (1973).

4.13. Recalculation of the pyroxene formulae

In order to make an estimate of the activity of jadeite component in an omphacite, it is first necessary to re-cast the pyroxene analysis in terms of stoichiometric end-members on the basis of 6 oxygens per formula unit.

Essene and Fyfe (1967) discuss the normative recalculation of clinopyroxenes and conclude that the jadeite mole fraction (X_{Jd}^{Cpx}) may be calculated in two distinct ways:

$$1. X_{Jd} = X_{Na} - X_{Fe^{3+}} \quad (-Cr)$$

assuming all Na is in Acmite or Jadeite end members. This method can only be used when a chemical determination of Fe^{3+} is available, as in Mäkinen & Howie (1972), and cannot be directly applied to microprobe data.

$$2. X_{Jd} = X_{Al}^{VI} - X_{Al}^{IV} \quad (\text{where } Si + Al^{IV} = 2.0)$$

assuming all Al is in Ca-Tschermaks molecule and jadeite. This method is very sensitive to the accuracy of the Si determination, which is often suspect in microprobe analysis (especially E.D.S. probe analyses).

Both methods suffer from defects, and neglect the influence of small quantities of Ti.

Partly to remedy this state of affairs, Powell & Powell (1974) have proposed a method of allocating cations which is independent of the Si analysis and which accommodates Ti in $CaTiAl_2O_6$.

It is this method of recalculation that has been used in this present study to determine the mole fractions of jadeite in microprobed omphacites (see Table 6).

Insofar as the Powell & Powell method is designed to assign Fe^{3+} and Fe^{2+} in microprobe analyses, there are various inconsistencies when one is dealing with pyroxenes with detectable amounts of both Ca-Tschermaks and acmite component. Since the Si analysis is ignored,

no Ca-Tschermaks molecule is assigned if acmite is also assigned, no matter how great the shortfall of Si atoms from 2.00 per 6 oxygens. This problem does not affect the validity of the derived jadeite mole fraction, as can be seen from a comparison of Mäkinen and Howie's (1972) wet chemical analyses of Groix pyroxenes with the present author's data (see Table 6).

If one uses the criterion $X_{Jd} = X_{Na} - X_{Fe^{3+}}$, all Mäkinen and Howie's pyroxenes fall in the range $X_{Jd} = 0.268 - 0.372$, whilst the present author's pyroxenes (from only two rocks) range from $X_{Jd} = 0.28 - 0.345$. Acmite contents are also comparable, and the average $Fe_2O_3/Fe_2O_3 + FeO$ ratio from published analyses of Groix omphacites (0.64) is consistent with the Fe^{2+}/Fe^{3+} assignment achieved by the Powell and Powell recalculation on the present author's analyses.

The errors implicit in recalculating X_{Jd} from probe data can be assessed by comparing the different X_{Jd} 's obtained by different methods. For omphacites with low Ca Tschermak's molecule, where $Si = 2.0 \pm 0.02$, and $Cr, Ti < 0.01$, the agreement of Powell and Powell's recalculation with that of Ryburn, Raheim and Green (1976) is to ± 0.04 mole fraction jadeite. The best agreement is between the $X_{Jd} = X_{Na} - X_{Fe^{3+}}$ method and Powell and Powell's method applied to wet chemically analysed pyroxenes ($\pm 0.01 X_{Jd}$).

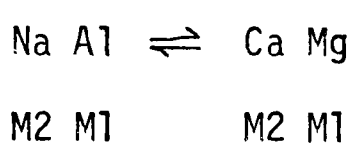
Uncertainties of the order ± 0.04 mole fraction jadeite would contribute 0.4 Kbar error to pressure estimates made at constant temperature.

4.14. Activity-composition relationship for jadeite in multi-component solid solution

After equilibrium temperatures have been estimated in a garnet + omphacite + albite + quartz assemblage such as PC10, and the X_{Jd} 's have been calculated as shown above, the problem of solving equation (1) for pressure depends solely on choice of mixing models.

Ideal 'molecular' mixing model

In this case, there is the maximum of short-range order and charge balance is conserved on the scale of the unit cell. Although there is random mixing on both of the two pyroxene octahedral sites, M1 and M2, substitutions on one are coupled to substitutions on the other, i.e.



Omphacites with $\text{Na} + \text{Ca} \approx 0.5$ and $\text{Al}/(\text{Al} + \text{Fe}^{3+}) > 0.6$ crystallize to a partially ordered structure with P2 symmetry at low temperature (Clark and Papike, 1966), so it may well be appropriate to apply this molecular mixing model to them.

$$a_{Jd}^{cpx} = X_{Jd}^{cpx} \dots\dots\dots 2)$$

This relationship is expressed on Fig. 7 as a straight line, with $\gamma = 1.0$.

Ideal ionic mixing model

At higher temperatures, short-range order begins to disappear, and ions are free to take up random juxtapositions between sites while mixing randomly on each separate site. Thus the activity of jadeite component can be expressed as the product of probabilities of finding Na on an M2 site and Al on an M1 site within one formula

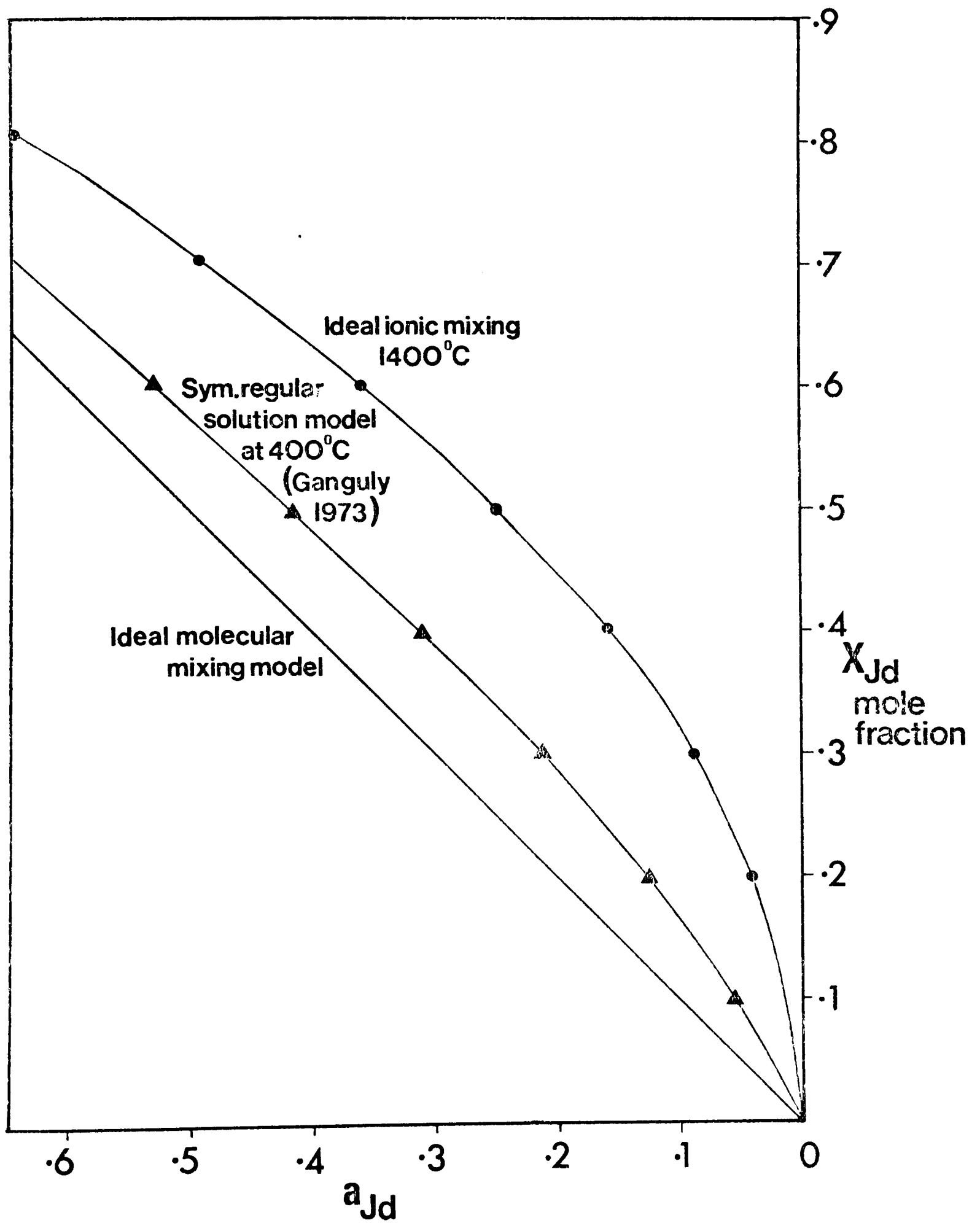


Fig. 4.7 Activity - composition relationship in system diopside - jadeite.

unit cell.

$$a_{Jd}^{cpx} = x_{Na}^{M2} \cdot x_{Al}^{M1} \cdot (x_{Si}^T)^2 \dots\dots\dots 3)$$

where x_{Na}^{M2} is the mole fraction of Na on M2 site, x_{Al}^{M1} is the mole fraction of Al on M1 and x_{Si}^T is the mole fraction of Si on tetrahedral sites. This is all assuming that Na and Al are perfectly ordered onto M2 and M1, though randomly mixed upon their respective sites.

This activity-composition relationship for jadeite is expressed in Fig. 7 as a curve corresponding to

$$a_{Jd}^{cpx} = (x_{Jd}^{cpx})^2$$

and would be most appropriate for C2/C symmetry clinopyroxenes at temperatures approaching their melting temperature ($>1100^{\circ}\text{C}$).

Symmetrical regular solution model (Ganguly, 1973)

Ganguly (1973) has developed an analytical expression for the activity of jadeite in natural multicomponent omphacitic pyroxenes, by assuming that a temperature dependent interaction between the jadeite and diopside/hedenbergite components is the source of a small negative deviation from ideality at low temperatures, i.e.

$$RT \ln \gamma_{Jd} = W_{Jd-Di/Hd} (X_{Di} + X_{Hd})(1-Jd)$$

where $W_{Jd-Di/Hd}$ is an adjustable interchange parameter, derived from a synthesis of experimental and natural paragenesis data:

$$W_{Jd-Di/Hd} = 1400 + 1.16 T (^{\circ}\text{C}) \text{ cal/mole}$$

Thus a_{Jd} can be extracted for any value of T by solving

$$\ln a_{Jd} = \ln x_{Jd} + \frac{W_{Jd-Di/Hd}(X_{Di} + X_{Hd})(1-Jd)}{RT}$$

This has been done for the omphacite analyses listed in Table 6, assuming the temperatures derived from the Raheim and Green geothermometer. The activity-composition relationship has been plotted for 400°C on Fig. 7, and is shown as a curve intermediate between the ideal molecular and ideal ionic mixing models.

Even if we are grossly in error in applying Ganguly's symmetrical regular solution model to the Groix pyroxenes, we know that the answer must lie between the two extreme cases represented by 'molecular' and 'ionic' ideal mixing. Assuming an equilibrium temperature of 410°C for the garnet/omphacite pair OMD-GTA in PC10, the ideal molecular mixing model ($\gamma = 1.0$) yields a maximum equilibrium pressure of 8.5 Kb whereas application of the ideal ionic mixing model results in pressures as low as 5.5 Kb. The preferred equilibrium pressure for this assemblage is 7.8 Kb, derived from an application of the Ganguly mixing model.

4.15. Comparison of Ganguly's model with natural paragenetic data

One way to check on the validity of the Ganguly mixing model for low temperature pyroxenes is to work backwards from a knowledge of the pressure and temperature of formation of a natural assemblage with P2 omphacites towards calculating jadeite activity. Jadeite activities having been estimated, it is then possible to draw iso-activity lines in P,T space and compare calculated activities with the activities derived from the extrapolation of natural samples.

This procedure has been attempted using the wet chemical analysis of an omphacite from Puerto Cabello (1059; Morgan, 1970) for which a single crystal structure refinement is available (Clark and Papike, 1966) showing P2 symmetry. Unfortunately, pressure estimates for this assemblage are very approximate, since no plagioclase is present. Assuming an ideal molecular mixing ($\gamma = 1$), the 1059 pyroxene

could have crystallized at 11.95 Kb in equilibrium with albite and quartz.

For the Ward Creek metabasalt (type II Coleman and Lee, 1963) 15CZ60 a minimum pressure can be fixed by the location of the aragonite/calcite transition, and a maximum pressure by the constraint $\gamma < 1.0$. This argument would also hold for the type III metabasalt 15CZ59 used by Ganguly (1973), but unfortunately no aragonite is present in the assemblage, so the minimum pressure constraint is more uncertain.

TABLE 7

	T°C	P _{min} [*]	P _{max}	X _{Jd}	a _{Jd}	Cal/mole Rtlnγ _{Jd}	Cal/mole W ₁₂
15CZ60	175 ⁰	4.8	6.3	0.389	0.26	-358	-2770 max
15CZ59	292 ⁰ ± 22	6.4	7.5	0.371	0.253	-429	-1312 max
1059	525	10	11.95	0.49	0.29	-800	-3337 max
	575	10.7**		0.49	0.42	-200	-850

Temperature estimates as in Table 5, except for 15CZ60, on aragonite-bearing metabasalts from Ward Creek. Temperature estimate derived from aragonite-lawsonite ¹⁸O partition. Taylor and Coleman (1968).

* Minimum pressures derived from aragonite-calcite transition of Boettcher & Wyllie, 1968a. For 1059, minimum pressure is derived from a_{Jd} = 0.37, assuming ideal ionic mixing: $a_{Jd} = X_{Na}^{M2} \cdot X_{Al}^{M1} (X_{Si}^T)^2$

** Preferred equilibrium pressure, assuming quartz and albite in

It is clear from the data presented in Table 7 that equilibrium pressures for low temperatures pyroxenes are not known with sufficient precision to enable us to distinguish between an ideal molecular mixing model and one entailing a small negative deviation from ideality. Even if we manage, as in the case of 15CZ60, to constrain the pressure between 4.8 and 6.3 Kb for 175°C, this does not help us choose an optimum value for a_{Jd} which is consistent with other pyroxenes equilibrated at different pressures and temperatures.

However, taking all three pyroxenes in Table 7 together, it is possible to reconcile their mixing behaviour (X_{Jd} vs a_{Jd} relationship) and pressure/temperature estimates with an iso-activity line corresponding to $a_{Jd} = 0.29$ (see Fig. 6).

This line places an upper boundary on the jadeite activity estimates for all the Groix pyroxenes (0.206 - 0.297) and so provides a relative framework for comparison. Until more detailed work is done on the crystal structure and site occupancy of the Groix omphacites, we are not in a position to choose a better mixing model. As it stands, the Ganguly model fits the rather crude requirements of experimental and natural data, and the consequent uncertainties in the estimation of equilibrium pressure remain small in comparison to the errors derived from disequilibrium problems, geothermometry, albite breakdown determination, chemical analysis and the re-calculation of mole fractions.

The present author estimates a total error of ± 1.5 Kb on the equilibrium pressure for the Groix omphacite/garnet pairs, and no more than 0.5 Kb of that error is contributed by uncertainties in the mixing model chosen.

4.16. Estimation of fluid phase composition during metamorphism

So far, the discussion of metamorphic conditions has centred on anhydrous assemblages where the fluid phase, if it ever existed, did not influence the solid-solid equilibrium.

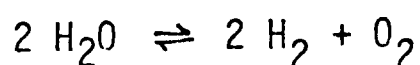
Triboulet (1974) discusses the role of water in the metamorphism of the glaucophane schists of Groix, and ascribes variations in paragenesis on the outcrop scale (say 10 m) to variable P_{H_2O} at constant temperature and P_{TOTAL} . This variation in P_{H_2O} could come about in many different ways.

- a) Original differences in degree of hydration of rocks before metamorphism.
- b) Hydrous phases could have buffered P_{H_2O} during prograde metamorphism in a closed system.
- c) P_{H_2O} could be controlled outside the system of interest, and water would be free to move throughout metamorphism and deformation according to P_{H_2O} gradients, tectonic fractures, etc.

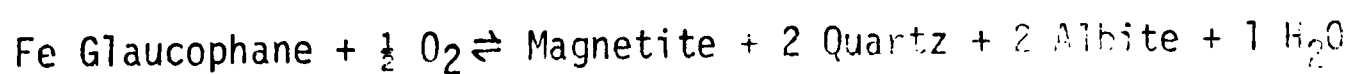
Unfortunately, it is very easy to confuse those differences in paragenesis due to varying water availability during prograde metamorphism with differences due to varying fluid composition.

It is perhaps more meaningful to talk in terms of oxygen partial pressure of the fluid phase (fO_2) during metamorphism, rather than trying to estimate the degree of water saturation or deficiency in a system.

During prograde metamorphism, fO_2 is controlled by the dissociation of water



and buffering of the fluid phase as various solid phases undergo redox reactions.



In addition, the presence of other fluid components, such as CO_2 and CH_4 , can act as a diluent, lowering the $f\text{O}_2$ for a particular pressure and temperature. With this in mind, it is possible to make various statements about the relative $f\text{O}_2$ during different episodes of metamorphism in the basic rocks of the Ile de Groix.

1. During prograde metamorphism of the zone I metabasics, $f\text{O}_2$ must have been high enough in some assemblages to de-stabilize lawsonite and favour the crystallization of iron-rich epidote (Pistacite = 30 mole %). On the other hand, $f\text{O}_2$ could never have risen so high as to de-stabilize ferroglaucophane. Assuming a maximum equilibrium temperature of 380°C at 8 kb for the non-garnet bearing ferroglaucophane metabasics (e.g. KB1), this means that $f\text{O}_2$ could never have exceeded the value defined by the Quartz-Fayalite-Magnetite buffer at that temperature, i.e. $10^{-31.6}$ bars (Hoffmann, 1972, see Fig. 8).
2. In the garnet-bearing glaucophane schists of zone II, epidotes are generally less iron-rich (Pistacite 20-25 mole %) especially when in association with lawsonite pseudomorphs. The inference that can be drawn is that $f\text{O}_2$ continued to fall as temperature increased and the rocks began to dehydrate. The absence of any primary ore phase in zone II, and the low $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratios of the basic rocks tends to support this argument.
3. Retrogression of the primary metabasic assemblages probably involved an increase in $f\text{O}_2$ with increasing temperature and decreasing load pressure. This is because magnetite, quartz and albite replaces ferroglaucophane in Zone I and ferric barroisite replaces glaucophane, omphacite and garnet in Zone II.

The first reaction has been calculated by Hoffmann (1972) to have a slope in $\log f\text{O}_2$ vs $1/T$ space so low that the curve is almost parallel to the temperature axis. Hence the magnetite + quartz +

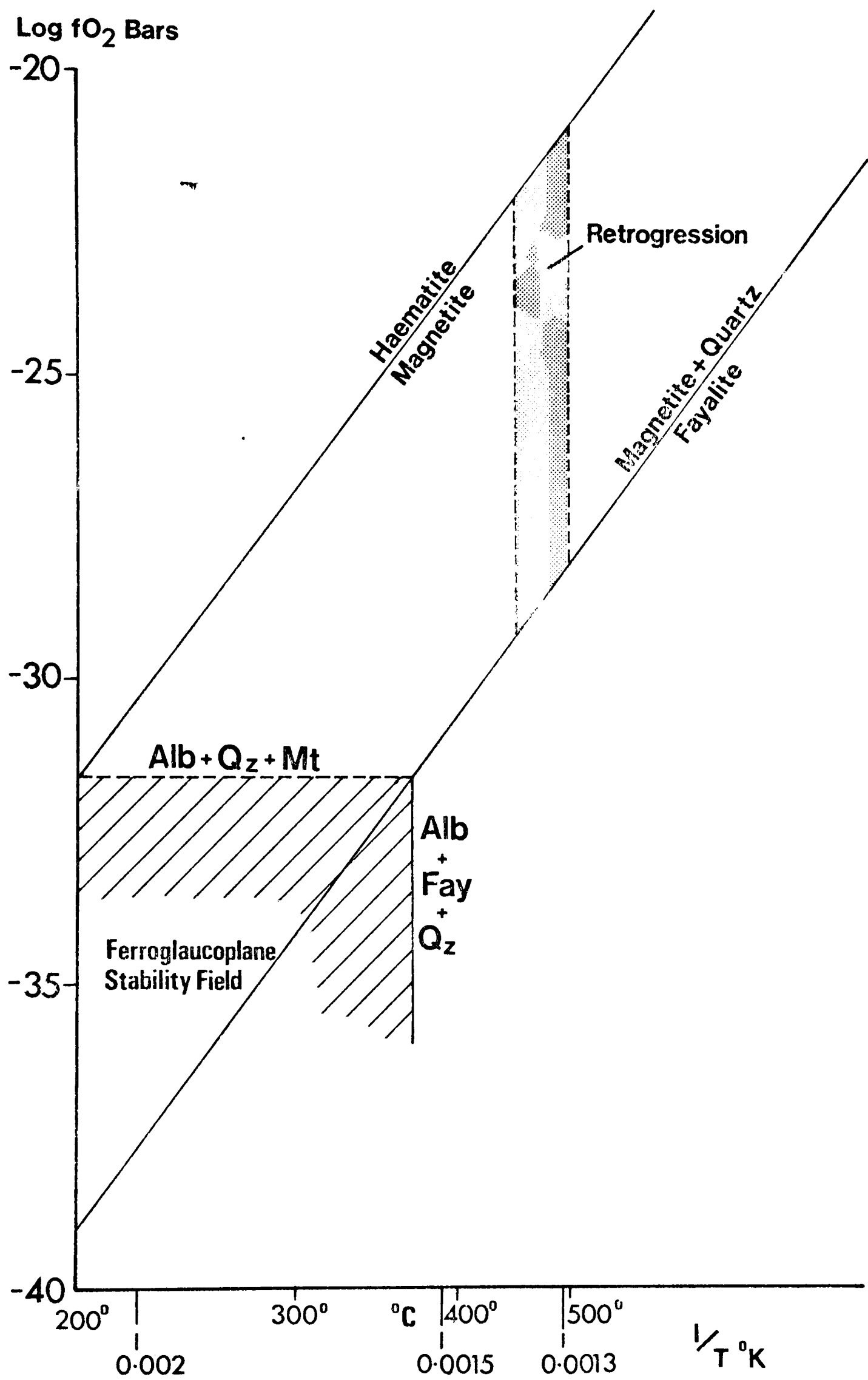
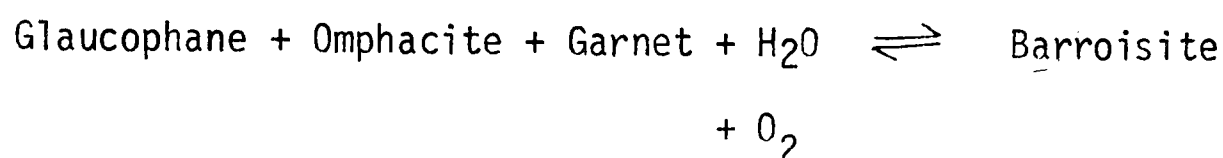


Fig. 4.8 T - f_{O₂} stability field for ferroglaucophane (from Hoffmann 1972).

albite stability field is constrained between the ferroglaucophane breakdown curve at low fO_2 , the haematite-magnetite buffer at high fO_2 and the Quartz-fayalite-magnetite buffer at low fO_2 and high temperature (see Fig. 8).

Similarly, with the second reaction;



the slope in $\log fO_2$ vs $1/T$ space is probably so steep as to be isothermal. In that case, any increase in fO_2 , as indicated by the ferroglaucophane reaction, must necessarily entail an increase in temperature.

Since haematite is very rare (c.f. spec LH14 of Mäkanjuola) in the retrogressive assemblages of Groix, it is probable that fO_2 during retrogression in most rocks never exceeded about 10^{-20} bars at a maximum temperature of 450-500°C. It is also quite likely that fO_2 during retrogression was everywhere in excess of values determined by the Quartz - Fayalite - Magnetite buffer (i.e. 10^{-30} - 10^{-28} bars in the range 450 - 500°C). The fO_2 -temperature field postulated for the retrogressive metamorphism (described in Chapter 2) is indicated in Fig. 9.

4.17. CO_2 content of the fluid phase attending metamorphism of the Zone I and II Groix metabasics

Ernst (1972) has estimated, using a $T-X_{CO_2}$ section at 7 kb computed from the sphene-rutile equilibrium of Schuiling and Wink (1967), that sphene + quartz assemblages lacking calcite are only stable at 400°C when $X_{CO_2} < 0.04$.

From this, he concludes that the composition of the fluid phase during the progressive metamorphism of the Sanbagawa blueschists was

influenced largely by dehydration reactions rather than decarbonation.

This result is consistent with the T - X_{CO_2} stability of lawsonite at 7 kb determined by Nitsch (1972), which places an upper limit of $X_{\text{CO}_2} = 0.02$ at 400°C for co-existing lawsonite + quartz + CO_2 .

The implications of this experimental work for the X_{CO_2} in the fluid phase during the metamorphism of the metabasics on Groix are as follows:

1. Lawsonite pseudomorphs bearing rocks (e.g. present author's KB1) with high Ca-content, rutile and quartz, but no calcite, probably equilibrated with a fluid with $P_{\text{CO}_2} > 100$ bars since calcite bearing assemblages at the same grade lack pseudomorphs and contain sphene rather than rutile (e.g. present author's specimen PN1). Sphene in equilibrium with quartz, calcite and epidote can only occur at temperatures near 400°C when $P_{\text{CO}_2} < 100$ bars and $f\text{O}_2$ is high enough to destabilize lawsonite.

$$\text{If } P_{\text{FLUID}} = P_{\text{H}_2\text{O}} + P_{\text{CO}_2} = P_{\text{TOTAL}} = 8 \text{ kb,}$$

then X_{CO_2} must lie between 0.0125 and 0.02 for lawsonite pseudomorphs and rutile bearing metabasics in Zone I. It could be that small decreases in X_{CO_2} influenced the relative stability of lawsonite in zone I at constant temperature by changing the effective $f\text{O}_2$. At low X_{CO_2} (< 0.125), $f\text{O}_2$ might be high enough to destabilize lawsonite, whilst a more CO_2 -rich fluid would dilute the $P\text{O}_2$ without any change of oxygen buffers, thus enabling lawsonite to crystallize within its $T - X_{\text{CO}_2}$ stability range.

2. Under the metamorphic conditions prevailing in zone II on Groix, lawsonite was no longer stable (see Chapter 3). Rutile is present in a few calcite bearing metabasics (e.g. present author's TUD5), thus fixing a minimum X_{CO_2} of 0.025 ($P_{\text{CO}_2} = 200$ bars at 450°C) for those particular assemblages.

Other rutile bearing assemblages contain no calcite, and are

characterised by rather high water contents and lack of omphacite.

Sphene is restricted to rather dry assemblages, where it is often found included in omphacite in association with quartz and phengite mica. The generally low fluid content of these last mentioned rocks (e.g. present author's PC10) during prograde metamorphism is reflected in the low $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio of the whole rock, and the low maximum X_{CO_2} of 0.0125 - 0.025 derived from Schuilling and Wink's rutile-sphene transition in the range $400^\circ - 450^\circ\text{C}$.

3. During retrogression of the Groix metabasics, sphene grew at the expense of any primary rutile. In view of the ubiquitous presence of sphene in chlorite + albite greenstones (even those containing calcite like the present author's spec LIN-1) P_{CO_2} could not have exceeded 200 bars at 450°C , which under a load pressure of ca. 5 kb (assuming a geothermal gradient of $25-30^\circ\text{C}/\text{km}$ and a mean density of 2.9) would imply $X_{\text{CO}_2} < 0.04$.

If X_{CO_2} in the retrogressive fluids were higher than during the prograde metamorphism, it is interesting to speculate on the source of this extra CO_2 . It would seem that the source of CO_2 must lie outside the present outcrop on the Ile de Groix, since there are no unmetamorphosed limestones observed on the island. Whatever the source of the retrogressive fluids, it is important to note that they buffered the locally developed retrogressive assemblages (e.g. magnetite, chlorite, albite, barroisite, sphene, calcite, actinolite, clinozoisite) and not vice versa.

4.18. Further checks on the pressure, temperature and fluid phase composition during prograde metamorphism on Groix

The pressure/temperature estimates quoted so far by the present author are based on partly theoretical, partly empirical extrapolations from experimental data, and the assumption of equilibrium having been achieved in natural assemblages. Evidence has been presented to support the argument for equilibrium between the co-existing minerals of carefully selected assemblages, and the conclusions drawn about equilibrium conditions are subject to various tests.

1. The absence of kyanite from the aluminous schists of Groix would suggest equilibrium temperatures below the upper stability limit of pyrophyllite (ca. 460°C at 8 kb, Kerrick, 1968). This statement probably holds true even if pyrophyllite itself is not observed since the first appearance of Al-silicate in quartz-bearing rocks undergoing contact metamorphism can be shown to indicate temperatures above the breakdown of pure pyrophyllite (410 - 430°C at 2-4 kb; Kerrick, 1968).

However, Mäkinen and Howie (1972) and Kienast and Triboulet (1972) believe that kyanite was once developed during the peak of metamorphism on Groix (viz. shimmer aggregates of Chinner, 1967), and they use this piece of 'evidence' to argue for high temperatures during a 'pre-glaucophanization' event. The temperature minimum chosen by Kienast and Triboulet is based on the pyrophyllite breakdown curve of Velde and Komprobst (1969) which gives 470°-510°C as the minimum temperature for the stable co-existence of quartz and Al silicate above 4 kb when $P_{TOTAL} = P_{H_2O}$.

Despite the alleged presence of 'shimmer aggregates' (Mäkinen and Howie, 1972) of unknown origin in the aluminous schists of Groix, it would seem safer to assume that kyanite was never stable

rather than base temperature estimates on an equilibrium which is not observed and for which experimental data remains conflicting.

2. The high Si^{4+} content of phengites from the basic schists of the Ile de Groix is well in accord with the experimental work of Velde (1965), which predicts a high pressure/low temperature stability field for celadonite-rich micas (40-50 mole %) in equilibrium with microcline, biotite, muscovite and quartz.

Although microcline, biotite and muscovite are absent from any prograde, phengite bearing assemblage on Groix, it may still prove possible to estimate an equilibrium temperature at constant pressure if the phengite is saturated in celadonite component. This situation can only arise in metabasic compositions, relatively poor in Al_2O_3 , where the phengite contains all the K_2O in the rock, but more Al_2O_3 and less $\text{FeO}_{\text{TOTAL}}$ and MgO than any other co-existing phase, in the presence of free quartz.

Hence, any phengite from an aluminous schist will always be undersaturated with celadonite component, and thus unreliable for producing temperature estimation.

It is significant that the present author has analysed phengites with the highest Si^{4+} contents (6.9 per formula unit) so far determined from the Ile de Groix, and that these phengites are in equilibrium with quartz, omphacite, albite and garnet (see Appendix IV on mineral chemistry: PC10 phengites).

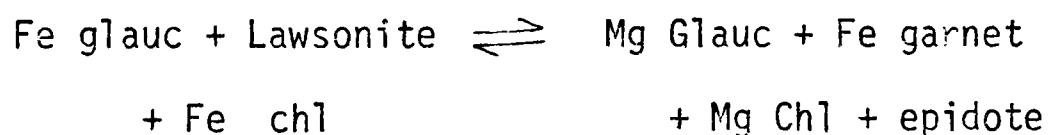
Assuming Velde's (1965) experimental data to be reliable and applicable to metabasic chemical systems without K-feldspar, muscovite or biotite, then the temperature of equilibration of a phengite with 6.9 atoms Si per formula unit (i.e. 45 mole % celadonite) would be 450°C at 8 kb $\text{P}_{\text{H}_2\text{O}}$.

All previous analyses of phengites from Groix give rather lower Si^{4+} contents than the PC10 example, but generally this can be

attributed to lack of saturation of celadonite component due to bulk chemical control. The data relevant to this problem is presented below in Table 8.

3. The stability of natural and synthetic ferroglaucothane has been determined by Hoffmann (1972) and this fits in well with the temperature estimates for the lawsonite-epidote transitional zone in the Ouégon schists of New Caledonia (Black, 1974). The relevance of a comparison with the zone I/zone II transition on Groix has already been discussed, so it is proposed to summarise the pertinent experimental data and postulated P/T conditions for Groix in Fig. 9.

Also on this diagram is marked the lawsonite-out reaction of Nitsch (1972), and a dashed line indicating the probable location of the 'garnet in metabasics' isograd:



Limits are placed on the P/T path of prograde metamorphism by using the jadeite activity and $K_D^{\text{ga-cpx}}$ data shown in Fig. 8, and assuming an average geothermal gradient of 15°C/km during burial.

The aragonite-calcite transition of Boettcher and Wyllie (1968a) is also indicated on Fig. 9, so as to provide a basis for comparison with the aragonite-bearing lawsonite glaucophane schist facies of the Franciscan.

The maximum temperature reached during prograde metamorphism is estimated by a consideration of the minimum $K_D^{\text{ga-cpx}}$ values (though adjacent mineral pairs such as PC10 GT1-OMPH1 may be disequilibrium) and the maximum probable error on the preferred temperature value ($400^\circ \pm 50^\circ\text{C}$). In addition, the maximum Si^{4+} content of phengite from metabasic assemblages, and the lack of kyanite in aluminous schists probably constrain this temperature to less than 450°C .

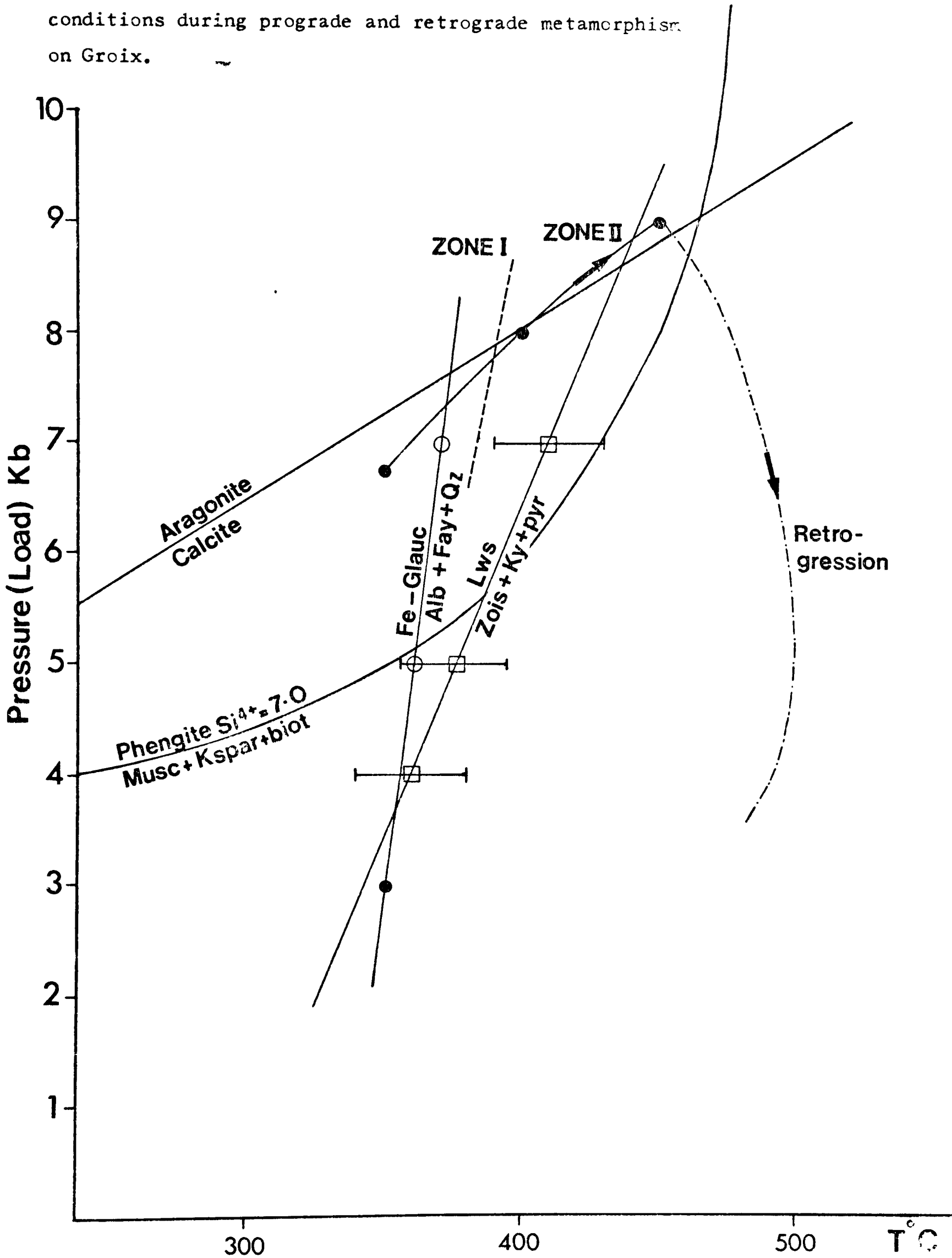
As far as the retrograde path (marked by dotted and dashed line)

TABLE 8 Si^{4+} Content of phengites

Phengite Sample	Si^{4+} per 24 O, (OH)	Paragenesis	Reference
IdG 2	6.38	Paragonite-bearing aluminous glaucophane schist Albite rich aluminous schists $\pm$ garnet *Chloritoid-bearing glaucophane schist	Velde 1967
3	6.44		
6	6.46*		
7	6.44		
9	6.40		
10	6.50		
Br. Is.	6.64	Chloritoid-bearing basic glaucophane schist	
Amer 4	6.454	Gneiss albitique a grenats	Triboulet 1974
Kerz 3	6.454	Micaschiste a grenats	
Sp 2	6.38	Gneiss albitique a grenats	
PdG 2	6.496	Gneiss albitique'a chloritoid	
Cd 52	6.53	Gneiss albitique a grenats	
LM2	6.542	Garnet bearing aluminous schists. $\pm$ glaucophane $\pm$ chloritoid	Makanjuola & Howie 1972
LH5	6.540		
LM5	6.535		
LM8	6.520		
LH10	6.521		
LH15	6.624		
PM6	6.528	Glaucophane-chloritoid aluminous schist	
LH13	6.595	Basic glaucophane-epidote garnet schist	
BM1	6.741	Magnesian schist. Talc-actinolite	
LH3	6.785	Glaucophanic eclogite	This study
LM3	6.614	Albite-epidote-glaucophane schist	
LH14	6.586	Barroisite bearing basic glaucophane schist	
PC10	6.9	Glaucophanic eclogite	
TUD5	6.68	Basic glaucophane-epidote-garnet schist	
KB1	6.54	Albite-epidote glaucophane schist	
PN1	6.74	Calcite bearing albite epidote glaucophane schist	

is concerned, very little can be fixed with any certainty, except to say that the increased 'A' site occupancy of the barroisite (ca 0.3 per 23 (O)) probably indicates temperatures in excess of the prograde conditions. Thermal modelling of the geothermal gradient during fast uplift (30 km in 40 m.y.) would seem to require a small input of heat during offloading, though this aspect of the 'retrogressive' history is more fully discussed in the chapter on geochronology.

Fig. 4.9 Summary diagram for proposed pressure/temperature conditions during prograde and retrograde metamorphism on Groix.



CHAPTER FIVE

GEOCHRONOLOGY OF THE ILE DE GROIX

5.1 INTRODUCTION

The data presented in this chapter represent the first attempt to use modern geochronological methods to date important tectono-thermal events in the evolution of the Groix blueschists. It should be stated from the outset that the present author has made no a priori assumptions about the nature or duration of metamorphic 'events', but simply has interpreted the mineral ages obtained in the light of the textural synthesis of Chapter 3 and the inferred pressure/temperature history outlined in Chapter 4 (Fig. 9).

Previous to the present study, ideas on the original age and subsequent metamorphic history of the Ile de Groix rocks have rested on a comparison with the structural history and lithologies of the S. Brittany mainland (Cogné 1960, Cogné et al. 1966). The present author believes that such analogies are not necessarily valid, and thus should be discarded as possibly misleading.

In particular, the idea that the Ile de Groix belongs to the margins of a belt that has suffered polyphase deformation and metamorphism due to the superimposition of a Hercynian orogenic cycle on an older, supposedly Cadomian (690 - 570 Myr B.P. C.J. Adams 1967 thesis) cycle, does not stand up to detailed textural and mineralogical investigation (see Chapter 3 section 5 . Objections to Cogné's model).

An important corollary of the polycyclic hypothesis is that the original depositional age of the sedimentary-volcanic formations on Groix must have been pre-Cadomian, and hence comparable in age to the late Pre-Cambrian Brioverian rocks of N. Brittany.

Recent whole rock Rb-Sr geochronological studies by Vidal (1972) show that the orthogneisses of S. Brittany (Massif de Moëlan, Landes de Lanvaux), considered by Cogné to be Cadomian, are in fact much younger

and define isochrons in the range 566 ± 8 to 443 ± 4 Myr. ($\lambda_{\text{Rb}} = 1.39 \times 10^{-11} \text{ yr}^{-1}$). According to Cogné et al. (1966) and Makanjuola (1967), the isoclinal fold style and amphibolite grade of metamorphism of the 'first orogenic cycle' on Groix is consistent with that observed in the Pouldu series (25 km away to the N.W. on the mainland) which is cut by the Moëlan orthogneiss. (see Figs. 1.1, 1.2).

Since the Moëlan orthogneiss near the contact with the Pouldu series has been dated by the Rb-Sr isochron method to be 485 ± 2 Myr old (Vidal, 1972), it now becomes conceivable that the Pouldu series was deposited some time in the Early Palaeozoic, and hence is not comparable with the Brioverian.

If the Pouldu series is equivalent in age to the rocks on Groix, then deformation and metamorphism must have occurred in both, though not necessarily simultaneously, between 485 and 290 Myr ago (this last date is fixed by the Rb-Sr mineral isochron from the Moëlan granite, Vidal, 1972).

Vidal (op. cit.) also quotes Rb-Sr whole rock data from Pouldu and Groix and concludes that the scatter of points is too great for the Pouldu series to permit the drawing of an isochron, though the four micaschists from Groix define a crude 'error-chron' of ca. 430 Myr. (This is further discussed in section 2.3 of this chapter).

This last piece of evidence would seem to suggest that the original depositional age for Groix could be as young as 430 Myr, so the time bracket for deformation, metamorphism and uplift becomes even tighter than for the Pouldu Series.

Using the data in this chapter, the present author would prefer to interpret the structural and metamorphic evolution of Groix as belonging to one, broadly Hercynian, sequence of events, especially since there is no evidence, geochronological or other, for Cadomian events in S. Brittany.

The results reported in this chapter come under three broad headings:

1. Sr isotope and trace element studies of the metabasic schists and some albitic schists of presumed igneous origin. It was hoped to place constraints on the maximum age and magma type of the basic volcanism on Groix (Sr isotope analyses by Dr. P.N. Taylor, Oslo 1975).
2. Conventional K/Ar dating of minerals and mineral pairs from rocks of known paragenesis and bulk chemistry (Carpenter and Civetta, 1976).
3. Rb-Sr dating of mineral pairs (isochron method) and single minerals (model age method) from the same rocks used in the K/Ar study (analyses by J-J Peucat, Rennes, 1975).

5.2. Strontium isotope and trace element studies on the metabasic schists, and some associated rocks, from the Ile de Groix

The problem of dating the original age of emplacement of the Groix metabasics, assuming that they were originally extruded rather than intruded, resolves itself into finding a suite of recognisably co-magmatic rocks that have suffered no Sr isotope exchange or changes in their Rb/Sr ratios to any significant extent after their formation.

With rocks that have probably suffered alteration upon eruption (i.e. spilitization), and which were subsequently metamorphosed at temperatures up to 450°C, it is obviously impossible to fulfil the requirements for rigorous Rb-Sr whole rock geochronology (Faure and Powell, 1972).

Nevertheless, if one is able to make assumptions about the initial $^{87}\text{Sr}/^{86}\text{Sr}$ of the magma, the degree and nature of alteration and the timing of metamorphism, it may be possible to place appropriate constraints on the maximum age of eruption.

During the course of the present study, 16 rocks from the Ile de Groix were analysed for $^{87}\text{Sr}/^{86}\text{Sr}$ (Dr. P.N. Taylor, Oslo) and Rb/Sr

ratios (M.S.N. Carpenter, Oxford).

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were determined on a Micromass 30 solid-source mass spectrometer, using standard ion-exchange column procedures for the separation of strontium (Pankhurst and O'Nions, 1973), and the Rb/Sr values were determined from pressed powder pellet XRF analyses, using the correction program due to Pankhurst (1969). (see Table 1)

Providing that all the metabasics on Groix represent lavas erupted within a short time-span, then the large range in $^{87}\text{Sr}/^{86}\text{Sr}$ observed can be explained in many different ways.

1. Original differences in initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio between contrasting magma types co-existing in the same tectonic environment.
2. Sr isotope exchange between sea-water and the basic rocks at the time of extrusion, if the eruptions were submarine. Also extensive sub-sea floor metamorphism soon after eruption due to circulating hydrothermal fluids (Spooner and Fyfe, 1973). If the water/rock ratio were high enough during hydrothermal alteration, this might enable quite efficient transfer of radiogenic ^{87}Sr into the metabasalts (Chapman, M.Sc. thesis, 1975).
3. Time integrated changes in Rb/Sr ratio during alteration and metamorphism would result in a scatter of present day $^{87}\text{Sr}/^{86}\text{Sr}$ ratios even if the original lavas had been co-magmatic.
4. During metamorphism, the metabasic schists might exchange Sr with the more radiogenic Sr in the surrounding pelitic schists. This process is only likely to be efficient if the Sr content of the pelitic schists is greater than that of the basic schists.

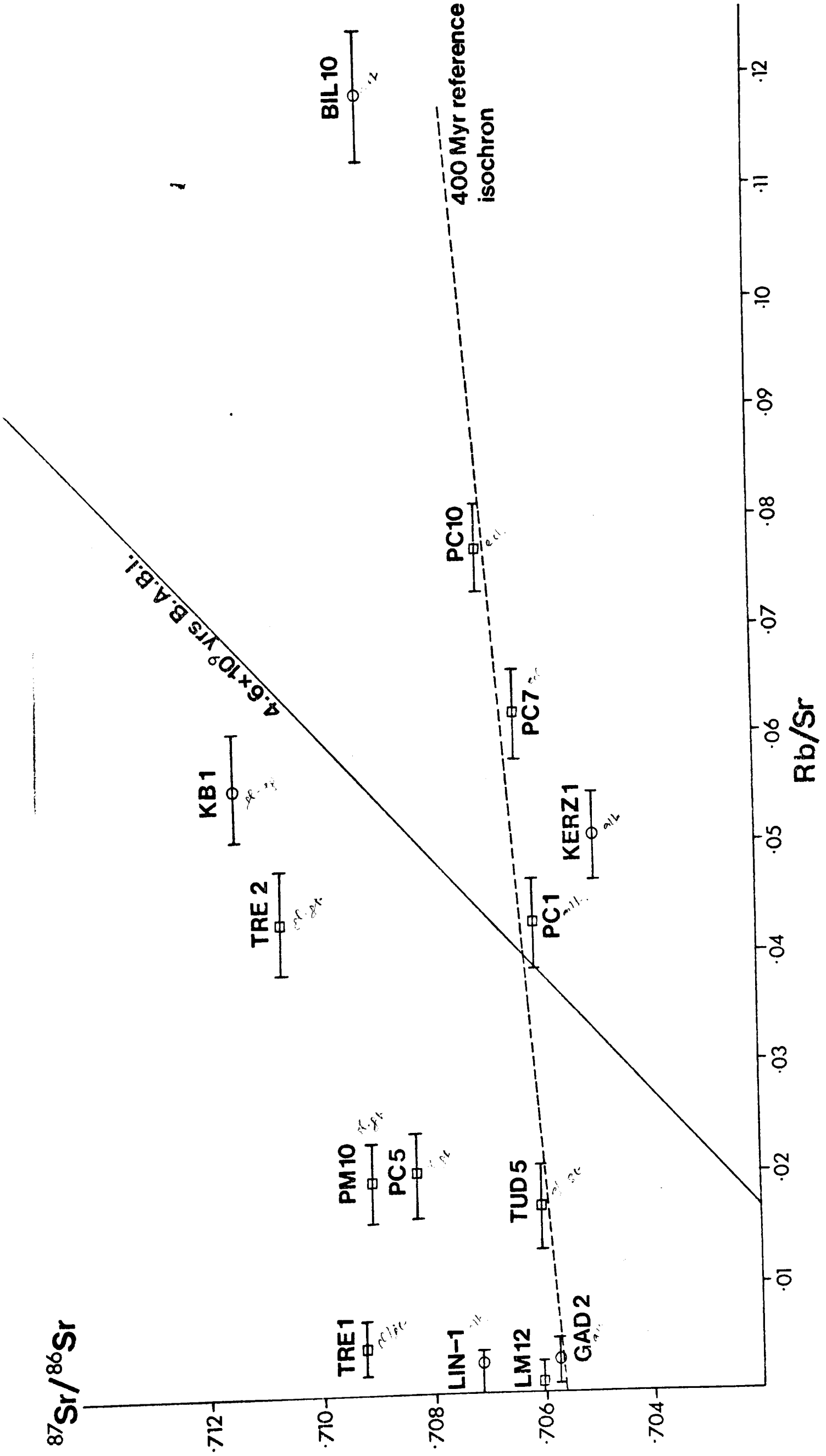
With all these factors in mind, it is not surprising that the data points of Table 1 fail to define any isochron (see Fig. 1). However, if one works on the principle that rocks with similar trace element characteristics are more likely to share the same $^{87}\text{Sr}/^{86}\text{Sr}$ initial

SAMPLE	$^{87}\text{Sr}/^{86}\text{Sr}$	present day	Rb/Sr	Rock type
PC 10	0.70704 ± 0.00016		0.077 ± 0.004*	Glaucophanic eclogites
PC 7	0.70642 ± 10		0.062 ± 0.004	
LM 12	0.70602 ± 14		0.001 ± $\frac{0.002}{0.001}$	
TUD 5	0.70601 ± 10		0.017 ± 0.004*	Garnet + Glaucophane + Epidote
PC 5	0.70836 ± 10		0.020 ± 0.004	
PM 10	0.70910 ± 6		0.019 ± 0.004	
TRE 1	0.70928 ± 7		0.004 ± 0.001	metabasics
TRE 2	0.71078 ± 4		0.043 ± 0.004	
KB 1	0.71153 ± 10		0.055 ± 0.004	Ferroglaucophane + chlorite + epidote
LIN-1	0.70710 ± 14		0.003 ± 0.001	
GAD-2	0.70576 ± 12		0.003 ± 0.001	+ Albite
PC 1	0.70610 ± 5		0.043 ± 0.004	
KERZ 1	0.70499 ± 12		0.051 ± 0.004	Greenstones
BIL 10	0.70917 ± 5		0.118 ± 0.006	
BIL 7	0.72065 ± 16		0.788 ± 0.026	Chlorite + albite amphibolite
TUD 6	0.72145 ± 7		0.819 ± 0.026	

The uncertainties listed above are all estimated two standard errors of the mean, except those marked * which were calculated from more than two replications. USGS standard BCR-1 was used to calibrate the XRF analyses of the metabasic rocks, and G-2 was used for the albitic micaschists. Standard values used for these USGS standards are those listed in Pankhurst and O'Nions (1973).

Measurements on the NBS SrCO_3 standard during the course of analysis in Oslo yielded an average $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.71037 ± 0.00005 .

Fig. 5.1 Rb/Sr versus $^{87}\text{Sr}/^{86}\text{Sr}$ plot for Groix metabasics



ratio, one can group the samples PC10, PC7, LM12, TUD5 and PC1 together as possibly co-magmatic.

If this is justifiable (c.f. Brooks et al., 1976) then the 5 samples quoted above would have shared an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7056 about 400 Myr. ago, assuming their Rb/Sr values did not change during the subsequent metamorphism.

On the other hand, the clustering of these five samples about the 400 Myr. reference isochron could simply be a fortuitous combination of sea-water alteration and open-system behaviour of Sr during metamorphism.

If this were so, then one would not expect to see any correlation between trace element groupings and Sr isotope composition. On the basis of the present author's trace element analyses, it would seem possible to distinguish several co-magmatic sub-groups amongst the metabasic rocks of Groix.

5.2.1. Trace element characterization of magma types

Rare earth element (REE) abundance patterns for the Groix metabasics disclose the presence of at least two, possibly three, contrasting magma types in the original volcanic pile (see Table 2, and Appendix II for analyses).

One group, which includes the samples PC10, PC7, LM12, TUD5 and PC1, is characterized by apparently flat REE patterns (enriched 10-30 times chondrite values, Nakamura 1974) across an MgO range of 4.12 - 12.69 wt.%, high Zr/Nb (> 25) and Y/Nb (> 3) ratios and Ni contents in excess of 60 ppm. (see Fig II. 4)

Another group is characterized by light REE enrichment (up to 60 X chondrite values), and included in this group are the samples KB1, LIN-1 and PC5. This group tends to lie off the main 'tholeiitic' trend on an AFM diagram towards more alkaline chemistry (see AFM

digram in Appendix II). Other characteristics include generally low Y/Nb (<2.0) and Zr/Nb (< 7.0) ratios and Ni contents less than 40 ppm for a MgO range of 2.65 - 6.15 wt.%. (see Fig II.5)

KERZ-1 is not included in this last group, although it has the highest light REE enrichment and lowest Y/Nb (0.8) ratio, since it can be distinguished apart by its very high Ni (182 ppm) and MgO (7.92 wt %) content.

5.2.2. Comparison of trace element geochemistry of metabasics with Sr isotope data

In Table 2 are presented the $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios for the metabasics calculated for a model age of 400 Myr. This model age is chosen because it corresponds with the reference isochrons shown in Figs. 1 and 3, but there are no rigorous grounds for believing the eruption age is anywhere near this estimate.

It can be seen from an inspection of Table 2 that there is a broad correspondence between the model $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio and the trace-element magma-type groupings as proposed in the preceding section. Choice of model age does not affect the validity of this broad correlation, but only changes the absolute position of the $^{87}\text{Sr}/^{86}\text{Sr}$ spread and causes a slight narrowing of the spread with increased age (i.e. Min. $^{87}\text{Sr}/^{86}\text{Sr}$ ratio at 400 Myr = 0.7043, $^{87}\text{Sr}/^{86}\text{Sr}$ max - $^{87}\text{Sr}/^{86}\text{Sr}$ min = 0.00638. Min $^{87}\text{Sr}/^{86}\text{Sr}$ ratio at 500 Myr = 0.704, $^{87}\text{Sr}/^{86}\text{Sr}$ max - $^{87}\text{Sr}/^{86}\text{Sr}$ min = 0.00625).

Evidence is presented in Fig. 2 to suggest that contamination of ^{87}Sr from sea-water or pelitic sediments cannot alone explain the large spread in $^{87}\text{Sr}/^{86}\text{Sr}$ observed for the metabasics.

If radiogenic ^{87}Sr had been introduced into the metabasics either during sea-water alteration or exchange with pelitic wall-rock, one would expect to see a negative correlation of Sr content with $^{87}\text{Sr}/^{86}\text{Sr}$

initial ratio. This is because the proposed contaminations contain less Sr than the basic rocks involved, and any process of exchange or assimilation might be expected to equalize Sr between starting material and contaminator.

Instead, a marked positive correlation is apparent in Fig. 2, suggesting that mixing of magmas or source material may have occurred before eruption.

Thus, on the basis of trace-element discriminant analysis (Pearce and Cann, 1973) and the model Sr isotope data presented in Table 2, it is proposed to distinguish three main original magma-types on Groix.

1. Ocean floor-like tholeiite-transitional basalts with flat REE patterns and high Zr/Nb (ca 30) ratios (Erlank and Kable, 1976). These rocks lie on a tholeiitic iron enrichment trend on an AFM diagram, probably largely controlled by olivine fractionation judging by the high Ni content (420 ppm) in the most 'primitive' sample PC1. $^{87}\text{Sr}/^{86}\text{Sr}$ initial = 0.7056 at 400 Myr ago, assuming no Rb/Sr changes during metamorphism. This magma-type also appears to show a distinct bimodality in TiO₂ (see Appendix Fig II.2.) and is more extensively developed in metabasics from the eastern half of the Ile de Groix (Zone II).
2. Ocean-island (within plate) type alkali basalts with light REE enrichment, low Zr/Nb (~ 7) and a large range in $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio from 0.707 - 0.711. It is possible that mixing of this magma type with a continental tholeiitic magma (with high $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio) accounts for the positive correlation of Sr with $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio, and the occurrence of certain samples (e.g. TRE 1, TRE 2 and BIL 10) with rather flat REE patterns but high $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio.
3. Within-plate, continental type basalt, highly enriched in light REE and many incompatible elements, but having low $^{87}\text{Sr}/^{86}\text{Sr}$ initial

Fig. 5.2 $^{87}\text{Sr}/^{86}\text{Sr}$ initial versus Sr ppm, assuming model age of 400 Myr.

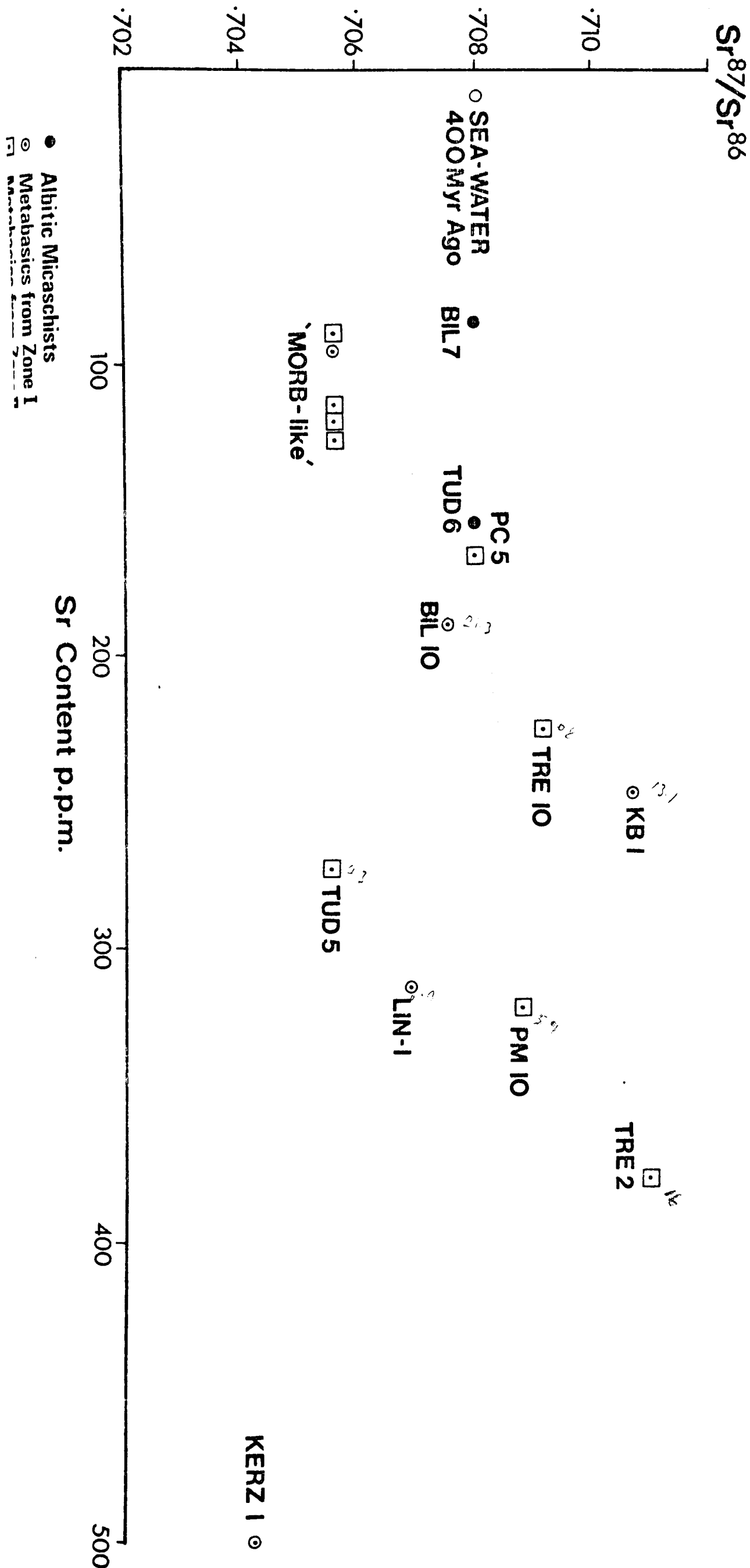


TABLE 2

	$^{87}\text{Sr}/^{86}\text{Sr}_i^{\dagger}$	Ce/Nd [*]	Nd/Y [*]
TRE 1	.70910	1.45	0.366
GAS 2	.70558	1.80	0.401
LIN-1	.70692	0.85	0.653
TUD 5	.70563	1.36**	0.341
LM 12	.70563	1.53**	0.272
PC 5	.70800	1.77	0.403
PC 7	.70563	1.47**	0.346
KB 1	.71068	1.64	0.770
PC 10	.70563	1.53**	0.310
BIL 10	.7075	1.76	0.344
BIL 7	.7080	1.96	0.727
PC 1	.7056	1.28**	0.384
KERZ 1	.7043	1.97	0.893
TRE 2	.710	1.47	0.308
PM 10	.7088	1.48	0.390
Chondrites		<u>1.37</u>	<u>0.35</u>

[†] initial ratio 400 Myr ago, calculated using present day Rb/Sr and $\lambda = 1.39 \times 10^{-11}$ yr. 2σ error = ± 0.000075

$^{87}\text{Sr}/^{86}\text{Sr}$ NBS Standard = 0.71037 ± 5

^{*} XRF determination using powder pellets, USGS standard values as in Nakamura (1974).

^{**} MORB-like magma type.

ratio = 0.7043. Judging by the high Ni and MgO contents, the rock KERZ 1 must represent either a fairly 'primitive' liquid or cumulate.

5.2.3. Estimation of the maximum age of eruption of the Groix metabasics

Although the present author would seek to explain the observed spread in initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios as primarily the result of original differences in magma type, this does not help in assessing the true Rb/Sr history of the metabasics subsequent to eruption, and hence great uncertainties remain concerning the absolute age.

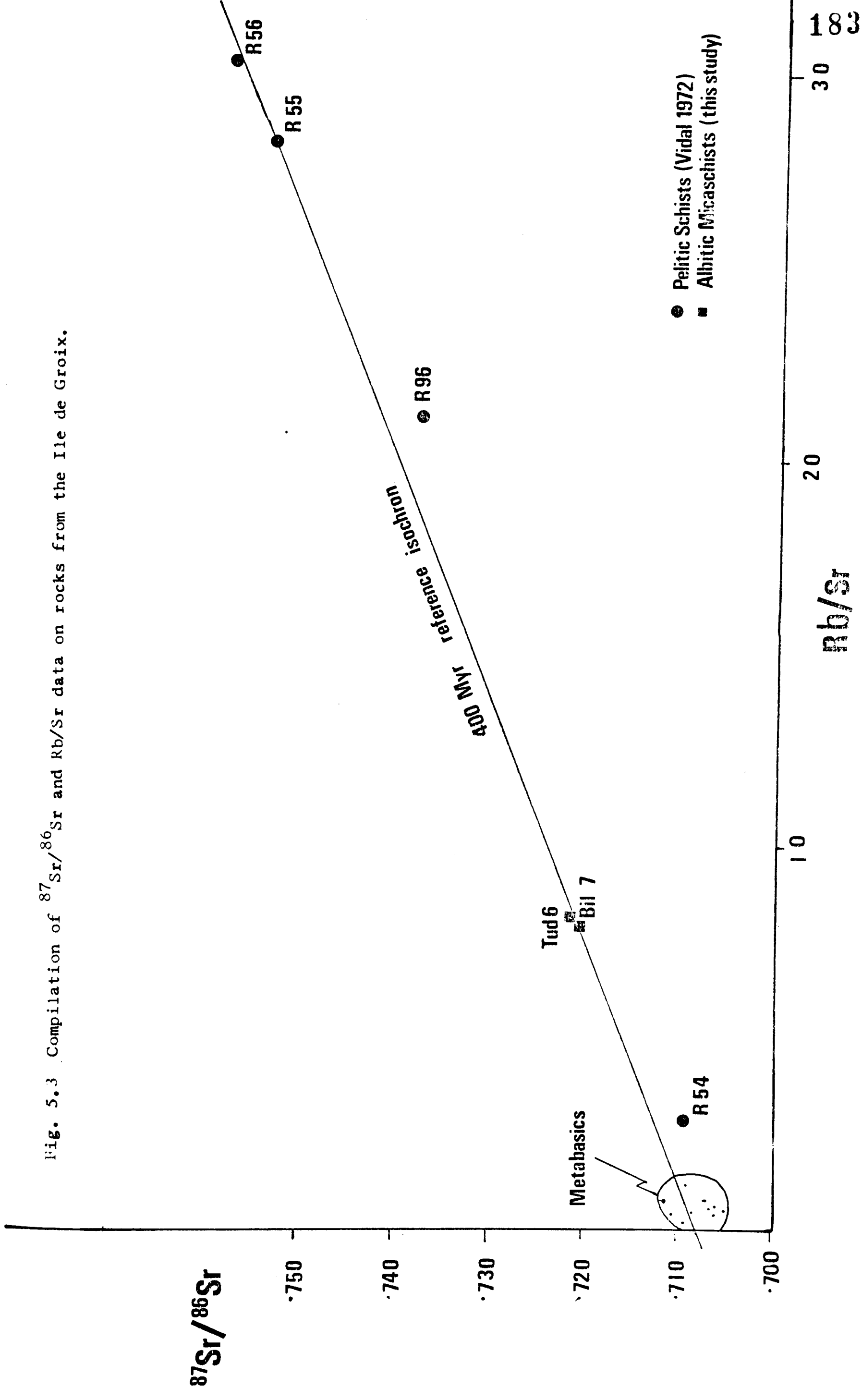
Fig. 3 shows all the whole rock Sr isotope data available for Groix, and at first sight would seem to indicate a major isotope homogenization event ca 400 Myr ago for rocks of widely different lithology. Unfortunately, the micaschists with high Rb/Sr (Vidal 1972) come from widely spaced outcrops, so it is not possible to assume an initial common isotopic composition of Sr for these rocks. In addition, sample R54 has a mineralogy very similar to that of the metabasic glaucophane schists (e.g. TUD 5), so one would not expect it to fall on any isochron with pelitic metasediments (R96, R55, R56).

The samples TUD 6 and BIL 7 are of albitic micaschists, which have the composition of acid or potassic igneous rocks (possibly rhyolites and keratophyre/mica lamprophyres respectively). It is interesting to note that these two samples fall on the same isochron as two of the micaschists; a situation that would arise if all four samples had shared an $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio of 0.7080 400 Myr ago and had not changed Rb/Sr during metamorphism.

Since Rb is more likely to be mobile than Sr during the hydrous metamorphism of these Rb-rich rocks, Rb/Sr would have probably decreased during metamorphism.

Now, taking the metabasic rock KERZ 1 with the least radiogenic Sr

Fig. 5.3 Compilation of $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr data on rocks from the Ile de Groix.



composition, and assuming that the albitic micaschists suffered only Rb/Sr decrease during metamorphism, it is possible to fix a maximum age, $t_{\max}$, for the eruption of the volcanics.

$$\frac{{}^{87}\text{Sr}/{}^{86}\text{Sr}_p \text{ TUD 6} - {}^{87}\text{Sr}/{}^{86}\text{Sr}_p \text{ KERZ 1}}{{}^{87}\text{Rb}/{}^{86}\text{Sr} \text{ TUD 6} - {}^{87}\text{Rb}/{}^{86}\text{Sr} \text{ KERZ 1}} = 1.39 \times 10^{-11} \times t_{\max} \quad \dots\dots\dots (1)$$

This equation assumes an approximation of the radioactive decay law

$$\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}_p} = \frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}_{\text{initial}}} + \frac{{}^{87}\text{Rb}}{{}^{86}\text{Sr}} (e^{\lambda t} - 1) \quad \dots\dots\dots (2)$$

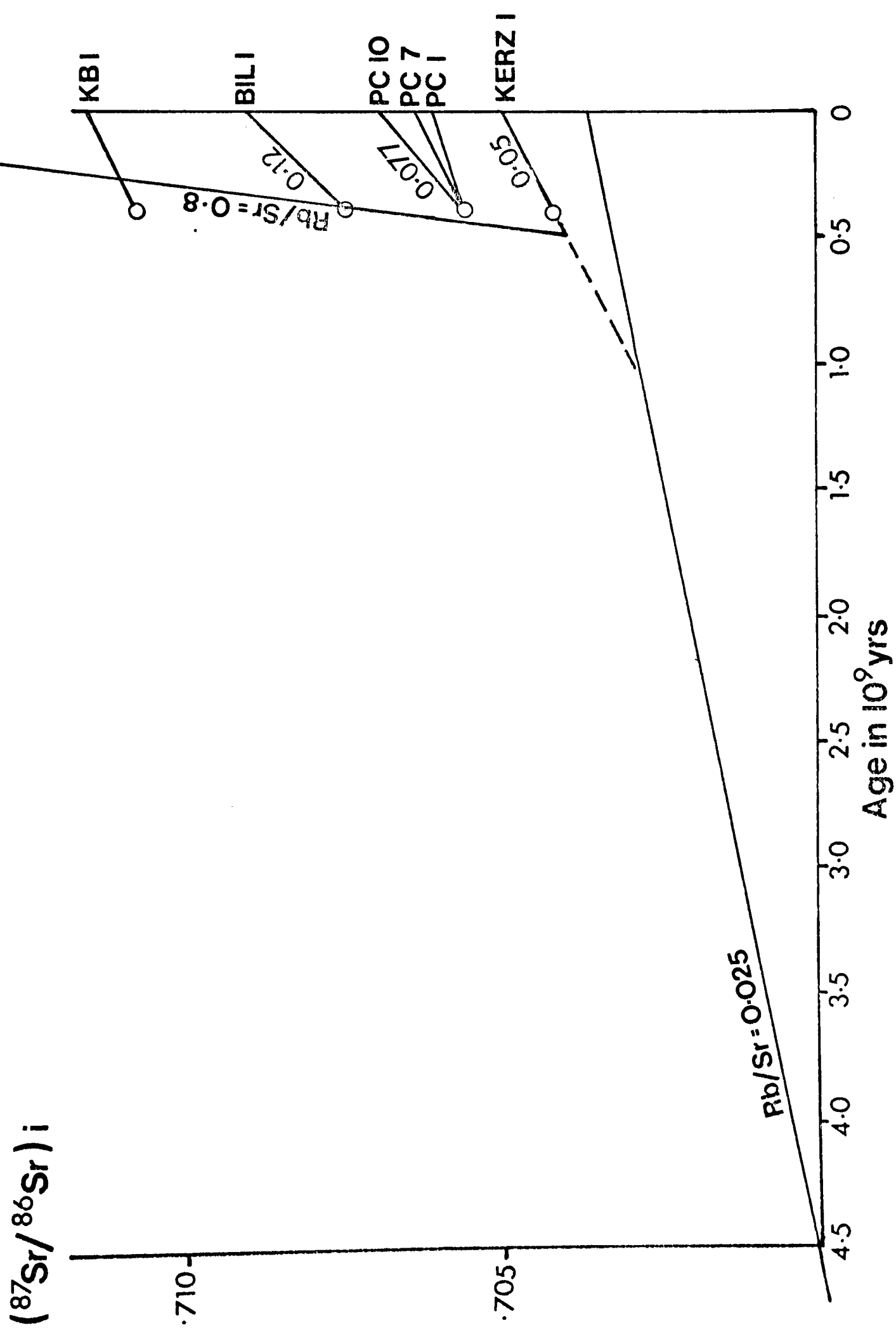
where ${}^{87}\text{Sr}/{}^{86}\text{Sr}_p$ is the ratio of these isotopes in the sample at the present day, ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ is the ratio t years ago, ${}^{87}\text{Rb}/{}^{86}\text{Sr} = \text{Rb/Sr} \div 2.8936$ and $(e^{\lambda t} - 1)$ is the slope of the isochron. For small values of λt , $e^{\lambda t} \approx \lambda t$ is a useful approximation and hence the simplification in equation (1). (λ_{Rb} is taken throughout to be $1.39 \times 10^{-11} \text{ yr}^{-1}$).

This operation is done graphically in Fig. 4, where the ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ evolution of TUD 6 and KERZ 1 is produced backwards in time, assuming constant, present-day measured Rb/Sr values. There is an intersection of the two lines at $t_{\max}$, 500 Myr ago, but this can only be regarded as a maximum age if we assume that KERZ 1 could never have had a higher ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ initial ratio than the associated 'keratophyres' or 'acid sheets'.

Changing the Rb/Sr of KERZ 1 during metamorphism between the limits 0.025 - 0.1 has very little effect on the value of $t_{\max}$, as long as the Rb/Sr of TUD 6 is kept constant at 0.8.

Although cases have been recorded of continental tholeiitic basalts with ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ initial ratios as high as 0.712 (Tasmania, Antarctica: quoted in Faure and Powell, 1972) and granites with ratios as low as 0.7032, (Salisbury Pluton, N. Carolina: also quoted in Faure and Powell, 1972),

Fig. 5.4 $^{87}\text{Sr}/^{86}\text{Sr}$ evolution diagram for Groix metabasites.



it is considered unlikely in the context of the metaigneous suite on Groix that these extreme cases could have occurred together in the same tectonic environment.

Furthermore, if one uses the micaschist R56 ($Rb/Sr = 3.08$) in equation (1) instead of TUD 6, t_{max} becomes 436 Myr. In this case, it is even more unlikely that the pelitic metasediments ever had a less radiogenic Sr composition than the interbedded basic volcanics.

In summary, the problem of the original age of the Groix volcanic-sedimentary pile must remain an open question, though none of the geochemical or Sr isotope evidence presented here is inconsistent with a date as young as 400 Myr.

5.3. K-Ar dating of metamorphic events in the Ile de Groix blueschists

In this section is presented an interpretation of K-Ar ages obtained on four glaucophanes, four phengites and one barroisite separated from aluminous metasediments, 'calcareous' banded epidote schists and massive basic glaucophane schists from the Ile de Groix. The data include two glaucophane-phengite pairs (PM6 and PC10) and an impure separate specially enriched in barroisite. (see Table 3).

Field relations and chemical mineralogy are best known for the glaucophanic eclogite PC10 (collected by the present author) which occurs as a boudin having its axis of elongation parallel with the dominant N-S mineral lineation in the S.E. of the island. This metabasic boudin lies in a matrix of chloritoid micaschist, and is apparently not an exotic block in the sense of Coleman and Lee (1963), since the encasing schist has an approximately co-facial paragenesis.

For the analyses of all minerals and rocks apart from PC 10 it is necessary to consult Mäkinen (1967) and Mäkinen and Howie (1972).

The pressure/temperature conditions during the metamorphism of the PC10 glaucophanic eclogite have been inferred in Chapter 4 of this study, and the estimate of 400°C at 8kb pressure is consistent with

TABLE 3

MINERAL ASSEMBLAGES OF WHOLE ROCK SPECIMENS
 LISTED IN ORDER OF DECREASING MODAL ABUNDANCE
WITH DATED MINERAL SEPARATES UNDERLINED

PM6: Phengite + Ferroglaucophane + chlórítoid + quartz +
 sphene + chlorite

Aluminous schist, possibly metatuffaceous sediment,
 from Stang er Marc'h near Pen Men (Zone I).

LM8: Phengite + quartz + garnet + chlorite + glaucophane.

Common pelitic lithology from Locmaria (Zone II).

LH11: Epidote + glaucophane + quartz + garnet + phengite.

'Calcareous' banded epidote schist, probably from
 margin of massive metabasic boudin, from Locmaria
 (Zone II).

LH13: Epidote + garnet + quartz + glaucophane + phengite.

'Calcareous' banded epidote schist (similar to sample
 R54 of Vidal, 1972) rich in mica. Probably from margin
 of metabasic body near Locmaria (Zone II).

PC10: Garnet + glaucophane/barroisite + omphacite + epidote +
phengite + albite + quartz + sphene.

Glaucophanic eclogite from Core of metabasic boudin,
 Pointe des Chats (Zone II).

Sample No	Mineral	^{40}Ar rad moles/g $\times 10^{-8}$	^{40}Ar rad %	K_2O %	$^{40}\text{Ar}/^{36}\text{Ar}$	$(^{40}\text{K}/^{36}\text{Ar}) \times 10^3$	Age (m.y.)
LM-8	Phengite	0.523	95.4	9.88 (± 0.04)	7234	329	328.3 (± 9.0)
LM-8	"	0.518	87.3	"	2470	105	326.0 (± 8.8)
LM-8	"	0.537	97.5	"	11863	544	336.3 (± 9.2)
PM-6	Phengite	0.544	84.7	10.28 (± 0.16)	1976	80.4	327.7 (± 10.3)
PM-6	"	0.543	95.1	"	6815	311	327.2 (± 10.3)
LH-13	Phengite	0.519	96.2	9.59 (± 0.19)	4851	215	334.6 (± 11.4)
LH-13	"	0.523	83.1	"	1859	71.9	338.9 (± 11.4)
PC-10	Phengite (< 100 mesh)	0.577	86.3	10.64 (± 0.09)	2230	88.9	339.3 (± 9.7)
PC-10	Phengite (> 100 mesh)	0.527	86.8	9.49 (± 0.09)	2311	92.3	342.7 (± 9.9)
PC-10	"	0.515	95.3	"	6451	288	335.5 (± 9.6)
PC-10	Glaucophane (80% pure) [†]	0.00422	34.2	0.0837 (± 0.0010)	461	8.31	313.2 (± 10.0)
PC-10	"	0.00437	58.3	"	765	23.0	320.7 (± 10.2)
PC-10	Glaucophane (70% pure) [†]	0.00593	17.8	0.118 (± 0.002)	361	3.23	313.2 (± 12.7)
PC-10	Omphacite + Gar net + Barroisite	0.00586	46.3	0.124 (± 0.001)	576	15.1	294.6 (± 8.2)
PM-6	Ferro-Glaucophane	0.000801	9.2	0.0151 (± 0.0005)	325	1.40	329 (± 33)**
LH-11	Glaucophane	0.000546	6.0	0.0107 (± 0.0005)	315	1.05	318 (± 32)**

$$\lambda_e = 0.585 \times 10^{-10} \text{ y}^{-1}; \lambda_\beta = 4.72 \times 10^{-10} \text{ y}^{-1}; {}^{40}\text{K}/\text{K} = 1.19 \times 10^{-4} \text{ atom/atom}$$

** The 1σ precision of the age determinations for PM-6 Glaucophane and LH-11 Glaucophane (with ^{40}Ar rad $< 10\%$) has been evaluated at 10%.

[†] Barroisite rims are the main impurity. Microprobe gives Barroisite rims $\text{K}_2\text{O} = 0.24\%$, Glaucophane Core $\text{K}_2\text{O} = 0.04\%$.

^{40}Ar was determined by isotope dilution mass spectrometry at the University of Naples, using a GD-150 Varian Mat mass spectrometer and the ^{38}Ar enriched Zurich spike. The calibration was checked by LP-6 USGS standard, a muscovite internal standard IMCT and Bern LP-6 Biotite.

K was determined in duplicate by flame atomic absorption with a Perkin Elmer 306 by the method of standard additions. Errors are maximum deviations from the mean.

The errors (1σ) for K-Ar dates have been evaluated by the Cox and Dalrymple formula. (Cox and Dalrymple, 1967)

A high pressure/low temperature metamorphic event 335 ± 20 Myr ago on Groix would post-date the intrusion of 'anatectic' granites on the mainland by about 40 Myr (Vidal, 1973), but would coincide with the intrusion of early Variscan granites throughout Brittany 340 Myr ago (Adams, D.Phil. Thesis, 1967).

This juxtaposition of blueschist with a region of active plutonism from 380 to 340 Myr ago in Brittany is analogous to that observed in the Mesozoic and Tertiary paired metamorphic belts of Japan and California. In the case of Groix, however, no 'median tectonic line' or suture is exposed, the blueschist 'belt' is restricted to one island, and the level of erosion of the 'hot belt' is so deep that the volcanic arc, if it ever existed, has been stripped off, revealing the underlying batholiths and migmatites.

As soon as subduction of the blueschists had ended, uplift was probably brought about by isostatic response to rapid erosion of the 'hot belt' Cordillera to the north. Thermal relaxation to a more normal (ca. $25-30^{\circ}\text{C}/\text{Km}$) geothermal gradient after subduction ceased would have produced a rise in temperature during off-loading, and a consequent retrogression of the blueschist facies minerals. If the uplift rate were fast enough, however, the blueschists of Groix would have reached the Earth's surface before the thermal relaxation 'wave' could erase much of blueschist assemblage.

The barroisites seem to have grown at the expense of glaucophane, omphacite and garnet (see Chapter 4) during a late rise in temperature that did not exceed the Ar-blocking temperature of glaucophane - otherwise the glaucophane dates would all have been reset at 295 Myr. This would account for the fact that the barroisite overgrowths on glaucophane are significantly younger by about 40 Myr than the high pressure/low temperature metamorphic assemblage.

Furthermore, the apparent slight discordance between the dates of

PC10 phengite and glaucophane can be explained by the influence of younger K-rich barroisite rims (probed by the present author: see Appendix III for analyses) in the impure PC10 glaucophane separates.

Recently, $^{39}\text{Ar}/^{40}\text{Ar}$ release age spectra (H. Maluski, unpublished work 1976, pers. comm.) have been obtained for several Groix glaucophanes which show a plateau age at 320 ± 10 Myr and no evidence for excess ^{40}Ar in the low temperature steps. This result is totally in accord with the conventional K/Ar ages reported here and in Carpenter and Civetta (1976), and the increased precision (± 10 Myr) quoted by Maluski allows a clear resolution of the barroisite age from the glaucophane. In addition, Maluski has observed a small low temperature step in the glaucophane release spectra, corresponding to an age of 300 Myr. The present author would choose to interpret this observation as evidence for the presence of a younger, retrogressive component (possibly barroisite).

The age of this late retrogression (ca. 295 Myr ago) corresponds well with uplift and cooling ages throughout the Hercynides, and is synchronous with the intrusion of younger Variscan granites (Adams, 1967; Vidal, 1973), a major break in sedimentation and the main phase of Hercynian folding (Sudetic phase: Westphalian in age).

5.4. Rb-Sr mineral isochron for the glaucophanic eclogite, and model ages for phengites from the Ile de Groix

All the phengite separates used for K-Ar dating (listed in Table 3), were also analysed for $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ p by J.-J. Peucat at Rennes University, France. Also, an epidote separate from the glaucophanic eclogite PC10 was selected as a possible Rb-free phase to define the $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio of the mineral system during metamorphism. The analytical results quoted by Peucat (1976, pers. comm.) are as follows:

TABLE 5

SAMPLE	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
PC10 Epidote	-	0.7055
PC10 Phengite	15.53	0.7808
PM6 Phengite	14.74	0.7767
LM8 Phengite	43.96	0.9243
LH13 Phengite	60.41	1.0259

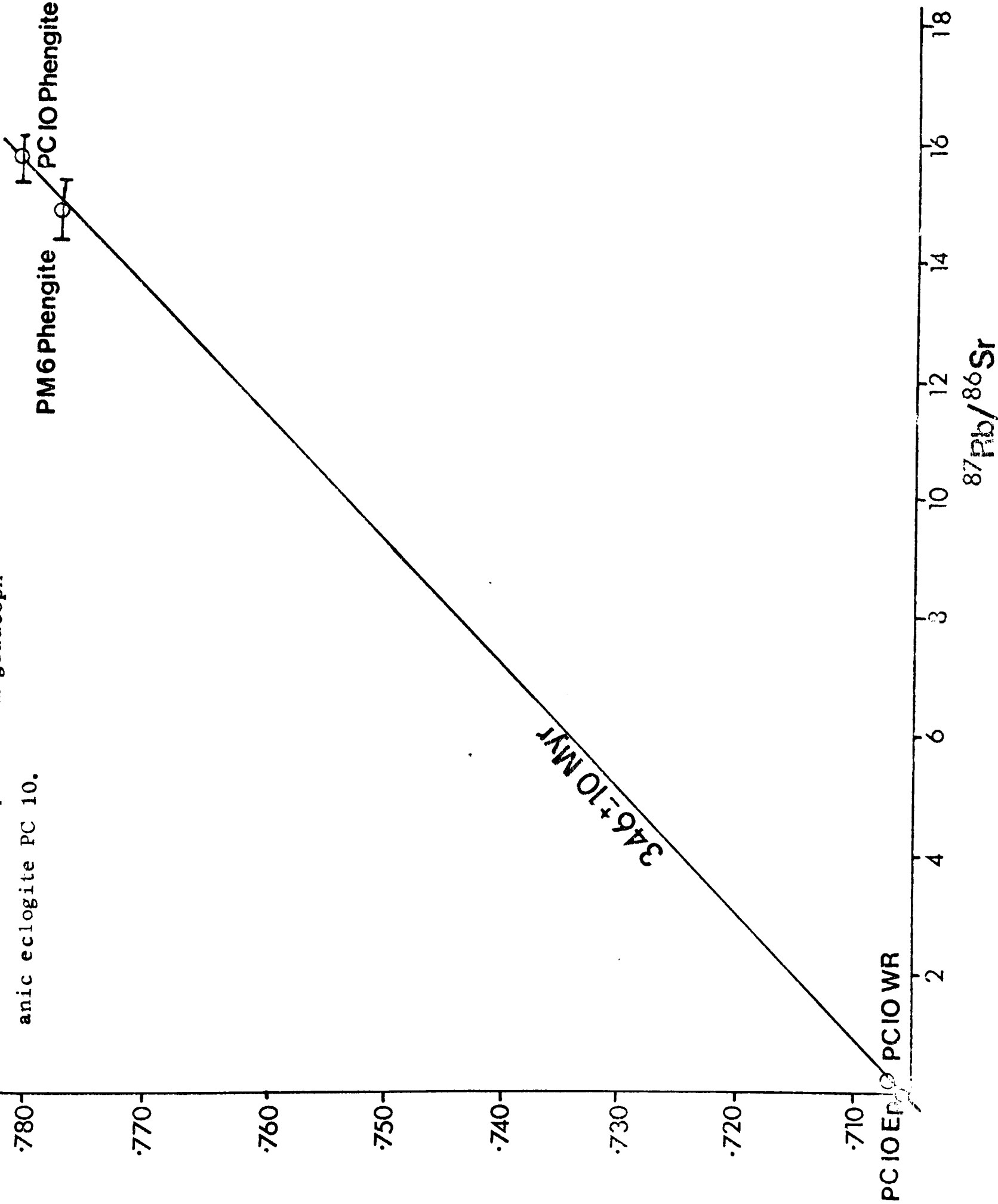
Errors on the isotope dilution determination of $^{87}\text{Rb}/^{86}\text{Sr}$ are estimated at $\pm 2\%$ (2σ) and the errors on the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio are estimated at 0.0005 (2σ). Repeated analyses of the Eimer and Amend SrCO_3 standard at the Rennes laboratory have given $^{87}\text{Sr}/^{86}\text{Sr} = 0.70816 \pm 0.00009$ (2σ) (Vidal, 1974).

The mineral pair Epidote - Phengite from PC10 defines an isochron, shown in Fig. 5, with a slope of 346.8 ± 10 Myr ($\lambda_{\text{Rb}} = 1.39 \times 10^{-11} \text{ yr}^{-1}$) that passes within error brackets of the PC10 whole rock data point (given in Table 1). This result strongly suggests that Sr isotopes were homogenized between all the constituent minerals of PC10 about 347 ± 10 Myr ago, and that diffusion of Rb and Sr ceased very shortly thereafter. The arguments used previously about K-Ar blocking temperatures exceeding mineral growth temperatures apply equally to the diffusion of Rb and Sr, and would lead us to suppose that the metamorphic event which brought about equilibration of Sr isotopic composition between minerals was reasonably short lived.

If metamorphism had proceeded for a very long time previous to

$^{87}\text{Sr}/^{86}\text{Sr}$

Fig. 5.5 Rb/Sr mineral isochron for
phengite-epidote pair from glaucoph-
anic eclogite PC 10.



340 Myr ago, then one would expect to see:

- a) excess ^{40}Ar generated from the decay of early formed K-rich minerals.
- b) discordance between K-Ar and Rb-Sr mineral dates from rocks of different grade.
- c) Rb-Sr 'mixed' ages due to the presence of an older, more radiogenic Sr component in zoned minerals and relicts.

In fact, there is no evidence for any of these phenomena, so a simple metamorphic peak age interpretation is preferred by the present author.

In the absence of analysed Rb-free minerals from PM6, LM8 and LH13, the interpretation of the phengite data is left in some doubt. One approach is to guess appropriate $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios for the lithologies concerned, assuming a short pre-metamorphic history. It is probably equally justifiable to assume concordant Rb-Sr mineral ages at 345 Myr, and thus find the necessary $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios for the mineral systems concerned.

TABLE 6

Model age for $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.705$		Model $^{87}\text{Sr}/^{86}\text{Sr}_i$ at 346 Myr
PM6	350 ± 10 Myr	0.7050
LM8	360 ± 10 Myr	0.7125
LH13	380 ± 10 Myr	0.7354

Since no whole rock Sr isotope data are available for rocks like PM6, on Groix we are in no position to judge what an appropriate $^{87}\text{Sr}/^{86}\text{Sr}_i$ should be.

However, for the aluminous schist LM8, we can assume that rocks

like Vidal's (1972) R55 and R56 should lie on a Rb-Sr mineral isochron including LM8 phengite. An isochron drawn through LM8 phengite and Vidal's whole rock micaschist data points for R55 and R56 yields an age of 350 ± 10 Myr, with an $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio of 0.7125.

This model age and initial ratio are consistent with both the K-Ar date from the same mineral and a short pre-metamorphic history (i.e. <100 Myr) assuming constant Rb/Sr values.

For LH13 phengite, the rather high model $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio at an age of 350 Myr is not unreasonable in view of the fact that the rock comes from the marginal zone of a metabasic boudin. It is possible that immediately post-eruptive alteration of the basic lavas produced very local high Rb/Sr patches (e.g. in glassy pillow rims) which evolved as closed systems for 50-100 Myr prior to metamorphism. There is also a strong indication from the zoned, rather 'sericitic' appearance of the phengite separated from LH13 (0.2 - 0.35 mm, K_2O wt.% = 9.39 - 9.78) that there may be disequilibrium problems effecting the Rb-Sr systematics.

5.5 Summary Statement

The Hercynian age proposed for the blueschist metamorphism on Groix is at variance with previous structural interpretations (Cogné et al. 1966) that have related early isoclinal fold phases to the Cadomian orogeny (690 - 570 Myr, Adams, 1967). The basic glaucophane schist assemblages have been considered by Cogné (1960) and MaKanjuola and Howie (1972) to have developed from pre-existing eclogites, possibly Cadomian in age. By contrast, it is considered very likely by the present author that both fold structures and metamorphic assemblages on Groix are Hercynian in age, and that the original lavas and sediments need be no older than 400-500 Myr.

A summary, Fig. 6, is presented which shows the relationship of the metamorphic events proposed by the present author for Groix with plutonic events, fold episodes and sedimentation history as seen on the mainland of Brittany.

The data for this Figure are extracted from Vidal (1972, 1973), Adams (1967), Auvray (1975), Vidal et al. (1971) and Michot and Deutsch (1970).

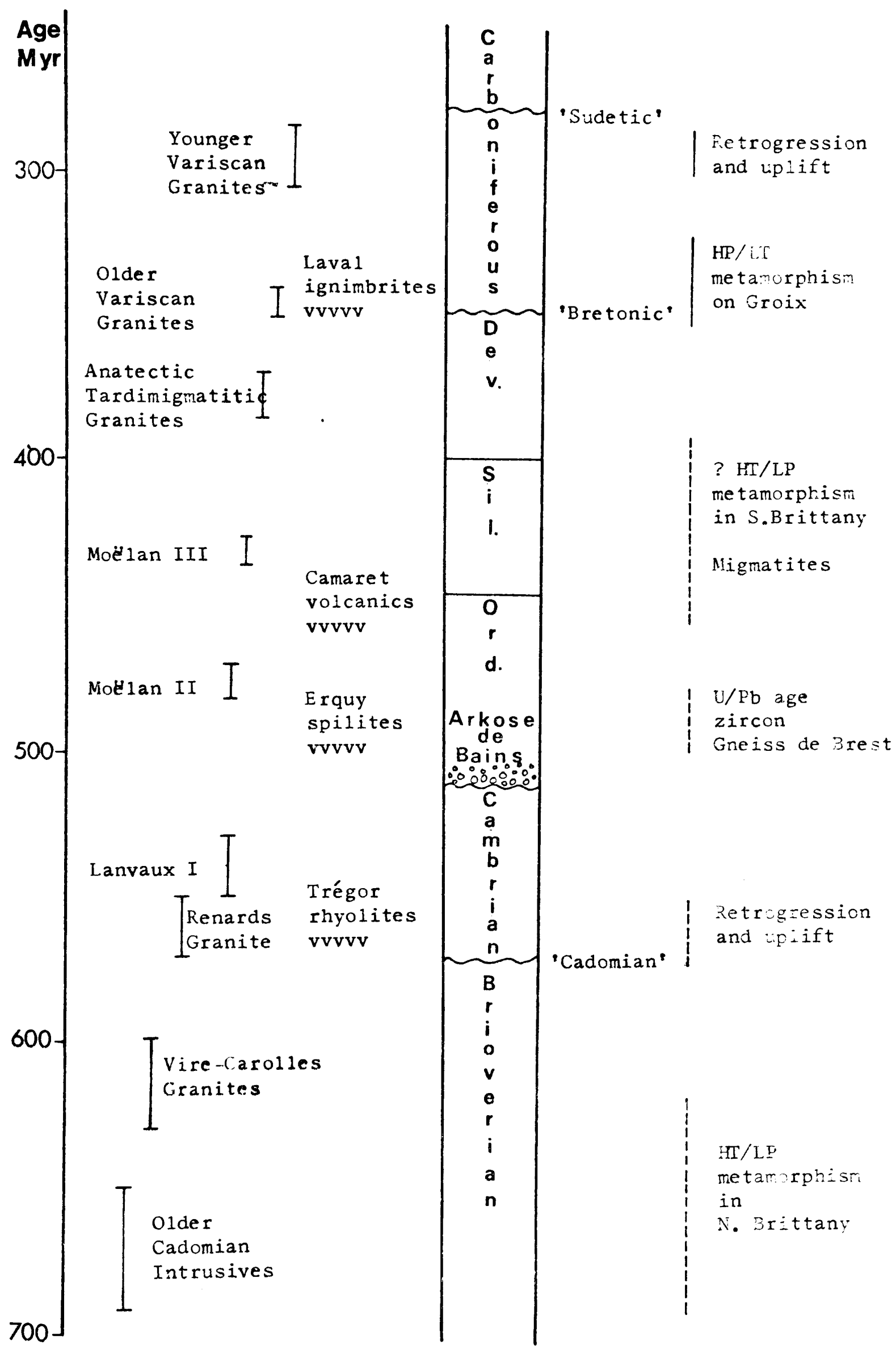


Fig. 5.6 Summary of plutonic, volcanic, sedimentation, folding and metamorphic events in Brittany.

CHAPTER 6

CONCLUSION

6.1 Introduction

The conclusions about the timing of high pressure/low temperature metamorphism on Groix presented in this study are to some extent dependent of the conditions of metamorphism inferred in Chapter 4. If the growth temperatures of the dated minerals were taken as 100°C higher, then it would be more difficult to maintain a short-lived metamorphic event model, but it would then become more difficult to explain the concordance of dates derived from different minerals and different methods.

However, the derivation of metamorphic conditions from mineral equilibria is independent of the assumptions necessary to interpret the geochronological data, and is subject to checking by comparison with other blueschist terrains. This comparison is only valid on the paragenetic level, and cannot be extended to geodynamic processes. The Ile de Groix may have very similar mineral parageneses to New Caledonia (as first noted by Briere in 1920) but this does not mean that similar tectono-thermal histories are implied.

Furthermore, tectono-thermal events in the evolution of the Groix blueschists need have no correlation with events on the South Brittany mainland metamorphic belt if both areas were separated by a plate boundary and thus experienced different histories up to 300 Myr ago. It is only at the very last stage of deformation and mineral growth (syn F_3 of the present author) that there is any similarity in the character of events between Groix and the mainland.

In the following section a hypothetical geological history for the Ile de Groix blueschists is reconstructed, using the evidence quoted in this study.

6.2 Synthetic geological history postulated for the Ile de Groix blueschists

Before their metamorphism, the glaucophane schists and associated rocks

of Groix were part of a thick volcanic-sedimentary pile laid down (or erupted) about 400 - 500 Myr ago.

This succession of strata was probably accumulated in a small marginal ocean basin, rather like the Japan Sea of today, but sufficiently near a continental margin to allow the influx of important clastic components and lavas of within-plate magma type affinities (Pearce and Cann, 1973). The range of volcanic rock types occurring together included spilitized ocean-floor tholeiites (especially near the bottom of the pile in zone II), alkaline basalts, minor keratophyres and lamphrophyres and rhyolites. It is very likely that the basic lavas suffered alteration and sub-sea floor metamorphism soon after eruption, leading to a differentiation of chemical composition between separate parts of pillows and flows. These small heterogeneities in composition (mainly Ca and Al) were later accentuated during metamorphism and deformation, and lead to the development of epidote-rich bands.

Associated sediments included carbonaceous shales, greywacke-type sandstones, volcanoclastic tuffs and iron-rich clays. It is possible that the proportion of sediments to lavas (ca. 3 : 1) increased towards the top of the succession (zone I).

The onset of metamorphism in the part of the volcanic/sedimentary pile now represented by the Ile de Groix began about 340 Myr ago, when the rocks were emplaced into a region of low (ca. 15°C/km) geothermal gradient by subduction at a trench system. Subduction may have been going on at the margins of the postulated small ocean basin for some time previous to 340 Myr ago - possibly since 500 - 450 Myr ago - but no record is now preserved of the subducted material. By analogy with Mesozoic Tertiary paired metamorphic belts (Miyashiro, 1973), it is possible that the 450 Myr migmatite 'ages' reported by Vidal (1973) for the Anticlinal de Cornouaille represent the peak of metamorphism in some 'hot belt'. Upward transfer of heat in the crust by granite plutonism may be the root cause of this thermal pulse, but whether the plutonism is related to some subduction process is not known.

The only evidence indirectly linking igneous activity with the postulated short-lived metamorphism on Groix is afforded by the outpouring of ignimbrites in the Laval basin of Central Brittany about 340 Myr ago, and the intrusion of many Older Variscan granites throughout Brittany.

Whatever the thermal history of the crust up to Hercynian times in the South Brittany area, it must have been totally different to that experienced by the 'marginal basin' crust now represented by Groix. This difference can be reconciled if a 'suture' or median tectonic line was active between Groix and the mainland metamorphic belt throughout the period 400 - 340 Myr ago. At the present day, the trace of this fault need not have any surface expression, and it may not even be in the place expected of it (i.e. between Groix and the mainland) due to late strike slip movements on the shear belts (see Fig. 1.2) and strong local differential uplift.

As soon as subduction ceased, and the Groix rocks had reached their maximum depth of burial (ca. 25 - 30 km), uplift would have commenced due to the

- a) rapid erosion of the overlying plutonic cordillera.
- b) strong positive buoyancy of the light (density = 2.9) metamorphic rocks, with respect to the surrounding mantle (density > 3.5).

There is no geodynamic reason why most of the uplift could not have taken place in 40 Myr, so that by 290 Myr ago the rocks of Groix stood at a sufficiently high level in the crust to experience brittle deformation and greenschist facies, water-rich, 'retrogression'.

The character of this late stage 'static' metamorphism is very similar to that recognized in metamorphic rocks throughout the Hercynides, and corresponds in age to the late Variscan granites and many mineral cooling ages (Adams, 1967; Vidal, 1972, 1973).

6.3 Problem of water supply during retrogression

The fast burial and subduction of the pre-metamorphic pile would be

conducive to the trapping of connate waters. Given reasonable estimates of subduction rate (3 cm/yr, based on present day spreading rates and a Benioff zone dip of 45°) the rock volume of Groix would have reached a depth of 25-30 km in less than 1 Myr, thus implying a very fast strain rate in view of the highly deformed nature of the rocks.

Under these circumstances, the wet buried rocks would be deformed in a ductile manner (isoclinal folding) whilst retaining locally their initial water contents, since porosity and permeability would be too low to allow transport of fluids over any great distance. This is because during progressive deformation under fast strain rates the shear strength of rocks would soon be exceeded and any voids containing fluids would close up. The initial water content of rocks would be taken up in hydrous metamorphic minerals, and any intergranular fluid is buffered by these minerals.

With high P/T (ca. 10 - 15°C/km), a prograde path of metamorphism might well lie along a line of equal molar volume for water (Norris and Henley, 1976), so there would be no tendency for hydraulic fracturing to occur in a rock system with no porosity and little permeability. During 'retrogression' and uplift, however, the style of deformation in the Groix blueschists was brittle, water was generally available in excess for the alteration reactions proposed in Chapters 3 and 4, and f_{O_2} increased above levels attained during prograde metamorphism.

It is proposed that during 'retrogression', the molar volume of water increased and trapped waters were released through the metamorphic pile from below by hydraulic fracturing. This hypothesis is supported by the occurrence of late cavities and veins, cutting all previous schistositities, filled with pegmatitic chlorite, albite and quartz (see Fig. 2.6).

6.4 Suggestions for the further testing of hypotheses given in this study

The equilibrium metamorphic conditions outlined in Chapter 4 could be refined by an increased knowledge of the space group symmetry of omphacites on

Groix and a more accurate understanding of the activity-composition relationship in the system diopside-jadeite at low temperatures.

There is also a need for oxygen isotope fractionation studies on mineral pairs such as:

Rutile - quartz	:	Zone I
Sphene - quartz	:	Zone II
Omphacite - quartz	:	Zone II
Garnet - quartz	:	Zone II
Magnetite - quartz	:	Retrogression

It should be possible, by selecting the appropriate paragenesis, to check the inferred metamorphic conditions given in Chapter 4, as long as isotopic equilibrium had been achieved and was not disturbed during subsequent events. There is certainly scope for fluid inclusion studies on Groix, since the present author has observed fluid inclusion, not only in the retrogressive assemblages, but also in quartz-rich fish net textured garnets from metabasic schists (e.g. present author's spec. PC5).

Experimental work could usefully be done on the sphene/rutile relative stabilities in the presence of lawsonite under low P_{CO_2} and low f_{O_2} at low temperatures, since this might support the idea that sphene could be developed post lawsonite breakdown in zone II on Groix, but before 'retrogression'.

Concerning the dating of metamorphic events on Groix, there is a need for more Rb-Sr mineral isochrons using omphacite, actinolite and epidote as Rb-free phases in order to resolve the age difference between 'prograde' and 'retrograde' mineral growth. The $^{39}Ar/^{40}Ar$ release age method could well be applied to mineral separates specially enriched in barroisite to discover if that mineral represents a younger, retrogressive component in mixed glaucophane ages. In addition, further work on the Rb/Sr systematics of carefully selected albitic micaschists might well provide tighter constraints on the original eruption/deposition age of the Groix rocks.

Finally, it might prove possible, using geophysical methods, to detect

some trace of the postulated suture between Groix and the rest of the Armorican Hercynides.

APPENDIX I
SAMPLE DESCRIPTIONS

LM12: 20 m east of the jetty at the east end of Locmaria plage.

Compact, massive 'beds' (15 - 20 cm thick) of fine grained glaucophanic eclogite, following intense planar S_1 parallel to regional schistosity (see Fig. 2.8). There is a pronounced streaking out of garnet augen (2 mm) parallel to glaucophane lineation ($118^\circ \rightarrow 135^\circ$), and a distinct lack of epidote-rich bands.

Instead, there are small (< 2 mm) discrete grains of epidote, often as cores to garnets, and highly altered patches of omphacite (not recognizable to the naked eye). The rock is very tough and splintery, of a slate blue aspect, and is sandwiched between epidote-banded schists which make up a unit at least 3-6 m thick traceable for 100 m inside the jetty. This type of outcrop is typical for the occurrence of pyroxene-bearing basic schists, as

figured in Mäkinen (1967), and Fig 2.8. About 500 g of the sample were selected for crushing.

PC10, PC7: Both extracted from a large loose block of glaucophanic eclogite found above the HWM on the west side of Por'h Morvil. Within 10 m of the block are in situ massive boudins of metabasic schists up to 8 m long and 1 m broad (see Fig. 3.3) and it is reasonable to suppose that the sampled block was derived from the core region of one of these boudins.

The rock is extremely tough and massive, with garnet porphyroblasts up to 4 mm in size, occasional wisps of pale green epidote and a strongly lineated felted mass of fine-grained blue amphibole. Omphacite laths up to 1.5 mm long and 0.2 mm broad are preserved in the 'pressure shadow' of large garnet porphyroblasts, and are lineated parallel to the glaucophane. White mica and sphene are important accessory minerals in this rock.

About 360 g of PC10 and 500 g of PC7 were selected for crushing. For a photomicrograph of PC10, see Fig. 3.11.

TUD5: Cote d'Heno, on west side of beach immediately west of harbour wall at Port Tudy. This locality is characterized by the occurrence of lawsonite pseudomorphs in epidote banded schists, and marked metasomatic effects near the margins of the metabasic horizons (viz. spec. PT2, an albitic micaschist 20 cm from contact with PT3, an epidote-rich glaucophane schist).

TUD5 itself was extracted from the centre of a large, possibly in situ, block embedded in the beach above the HWM. The rock is almost black on the weathered exterior, and may correspond with a band of black, smoothly polished fine grained rock exposed further around the headland to the west (cf. spec. TUD4, an epidote bearing glaucophane schist with calcite).

When broken, TUD5 exhibits hardly any banding at all, but abundant 4-5 mm porphyroblasts of garnet are apparent in a fine grained matrix of epidote and blue amphibole.

Over 1 kg of the sample was selected for crushing.

PC5: East side of Por'h Morvil, 2 m beneath the margin of a metabasic boudin (cf. spec. PM10) just above HWM. This rock is massive, tough and fine-grained, with a steel-blue weathering colour and large (5 mm) garnet porphyroblasts. Epidote is very scarce, and omphacite has never been found in this specimen. Over 1 kg of this rock was selected for crushing.

TRE1: North-west side of the Plage du Trec'h, just below HWM and within 2 m of TRE2. The rock is very fine grained, and light blue in colour when broken.

TRE2: Same locality as for TRE1. This outcrop is at the margin of the massive metabasic body sampled in TRE1, and is characterized by the occurrence of very large garnet porphyroblasts (6 mm) concentrated into an almost monomineralic seam (0.75 cm).

The rock contains epidote rich bands folded into flat lying isoclinal folds (see Fig. I.1) with rounded hinges. Glaucophane laths are aligned parallel to the b axis of the isoclinal folds, and are refolded by upright F_2 crenulations (trending 160°).

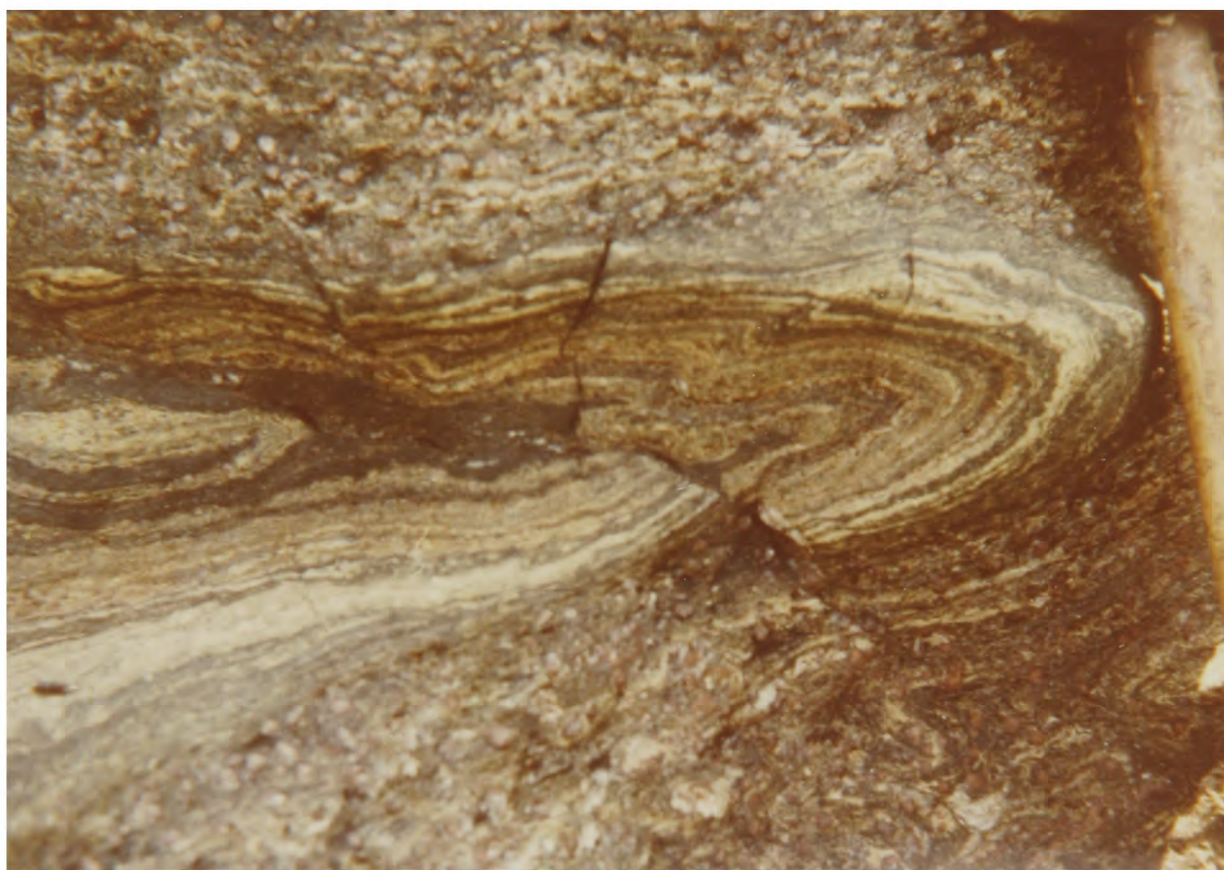


Fig. I.1

Epidote rich bands folded into flat-lying F_1 isoclinal folds. Note also large garnet porphyroblasts at top of photo. Locality: Plage du Trec'h, TRE2. Hammer is 3cm wide.

PM10: East side of Por'h Morvil, in a low cliff exposure 2 m above HWM.

This specimen comes from a band at least 30 cm long and 10 - 15 cm wide near the top of a large metabasic boudin (cf. spec. PC5). The rock is coarse-grained with 3 mm porphyroblasts of chloritoid visible to the naked eye, and very large pseudomorphs after lawsonite developed on weathered schistosity surfaces. Epidote forms large, streaky patches (not continuous bands) and lends a pale yellowish green, mottled appearance to the rock. Glaucophane is apparent as small, stubby, decussate laths. Small pinkish euhedra of garnet are also seen, though in minor quantities. For a photomicrograph, see Fig. 3.14.

KB1: West side of Baie des Cures, on cliff tops near a blowhole. The rock consists of stubby, crudely orientated laths of blue-black glaucophane (up to 6 x 2 mm) in a matrix of fine grained epidote. Lawsonite pseudomorphs weather out as rectangular shaped lumps, also crudely orientated parallel to the glaucophane lineation ($42^{\circ} \rightarrow 224$). In appearance, the rock is mottled dark-blue with a shimmer of white mica on the S_1 surfaces. Banding is not apparent in the specimens selected for analysis (500 g). The rock passes up into highly albitized quartz micaschists, then a massive chlorite + albite greenstone in the space of several metres.

For a photomicrograph, see Fig. 3.13.

PN1: 100 m north of the Pointe de St. Nicholas, from a cliff top exposure.

The rock is characterized by intense cm scale epidote banding, folded into open style F_2 isoclinal folds with axial planes dipping steeply S.W. Both glaucophane and epidote are granoblastic, fine-grained, and appear highly altered. Spots of brown-stained calcite are also very apparent, but no pseudomorphs after lawsonite are seen.

GAD1, GAD2: On a narrow ledge, 10 m down from the cliff tops 100 m east of the headland named Gadoeric. This locality is pictured in Fig. 2.4, and is characterized by the contact of a massive chlorite + albite greenstone (GAD2) with a highly sheared banded epidote schist (GAD1). The banded epidote

schist has abundant fine grained glaucophane needles showing a pronounced lineation parallel to the fold axes of the disrupted, flat lying isoclinal in the epidote bands. GAD2, the massive greenstone, was sampled only 2 m from GAD1, and has a cavernous, brown weathering aspect, with 5 mm blebby albites and fine grained dark green chlorite.

PC1: On the foreshore immediately west of the Pointe des Chats. This outcrop is pictured in Fig. 2.10, and a distinct gradation between crumbly, massive greenstones and compact glaucophane garnet schists is apparent in the photograph.

The rock itself is massive, with highly altered residues of garnet porphyroblasts in a matrix of fine grained chlorite and blebby albite.

LIN-1: From the east side of Port Melin, on a rocky knoll rising above the beach. This rock is brownish, cavernous weathering and very similar in aspect to all the other chlorite + albite greenstones (e.g. GAD2, KERZ1, TUD7).

KERZ1: At the western end of Locmaria Plage. The outcrop consists of a 5 m thick bank of massive chlorite + albite greenstone sandwiched between pelitic micaschists.

BIL3, BIL10, BIL7: In the Crique de Bileric, accessible by a narrow steep path from the cliff top. The outcrop for these specimens is pictured in Fig. 2.6, which shows a massive, fine-grained metabasic rock (BIL10) passing upwards into an albitic micaschist (BIL7) occupying a narrow band parallel to schistosity.

The rock BIL3 is coarse-grained, with rectangular simple twins of albite up to 6 mm long and abundant light green, radiating and decussate needles of actinolite (2 x 0.07 mm). Only 360 g of this rock were available for analysis, and a photomicrograph is presented as Fig. 2.11.

BIL10 has a compact, dense appearance, with very fine mm scale laminations of dark green and black material. Schistosity surfaces are scattered with planar orientated white mica; otherwise the constituent minerals are too fine grained to see with the naked eye. In Fig. 2.6, drusy

cavities within the rock are filled with pegmatitic quartz and albite.

BIL7 is not in direct contact with BIL10, since a narrow band of massive chlorite + albite greenstone intervenes. In appearance, BIL7 is not unlike the massive greenstones, except that white mica is much more abundant and there are a profusion of dark green ? chloritic laths.

TUD6, TUD7: 150 m east of the eastern harbour wall of Port Tudy.

A massive, highly weathered chlorite + albite greenstone horizon (12 - 18 cm thick), TUD7, is sandwiched between albitized micaschists with a strongly developed S_1 schistosity. The zone of albite-rich micaschists is only 1 m thick on each side of the greenstone, and it is from this narrow zone that TUD6 is taken. At distances greater than 1 m perpendicularly away from the greenstone contact, albite content in the micaschists falls off sharply.

APPENDIX II

Whole Rock Chemistry

The major element analyses quoted in Table II.1 were performed on a Phillips PW 1410 XRF Spectrometer at Oxford, using fusion beads prepared according to the method outlined by Norrish and Hutton (1969).

Estimates of error, the relevant machine settings used and a resume of the bead preparation specifications are given in Carter (1976).

Raw count data were corrected for inter-element adsorption and fluorescence effects using the coefficients given by Norrish and Hutton (1969) and the computer program "XRF" (written by Dr. S.W. Richardson) based on the calculation method of Harvey et al. (1973).

The standard values used for calibration were those reported by Flanagan (1973) for the USGS rocks PCC-1, DTS-1, W-1, BCR-1, AGV-1, GSP-1 and G2.

Na₂O was analysed by flame atomic absorption spectrometry on a Perkin-Elmer 306 machine at Oxford, using aliquots prepared from HF and HClO₄ dissolution of rock powders and NaCl standard solutions. Precision (maximum deviations from mean of replicate analyses) is estimated at $\pm 3\%$. Accuracy was checked against a solution made up from the USGS standard BCR-1 (Na₂O wt% recommended = 3.26). Results obtained fell within the quoted precision of the recommended value.

FeO was determined by Wilson's metavanadate back titration method, and the precision is estimated at $\pm 2.5\%$ (2σ).

'H₂O' was determined by the Penfield tube method for only 10 of the samples in Table II.1. Due to lack of prior heating to 110°C, and the probable partial release of CO₂ at 800°C, these analyses should be regarded with some caution.

The loss on fusion (20 min at 1000°C) is expressed as a percentage of the starting (wet) sample weight, assuming a constant dry flux weight of

2.00g, and ideally should correspond with total volatile content if all Fe was present as FeO. However, some weight loss on fusion is due to reduction of Fe_2O_3 and release of O_2 and some other losses may result from a dry flux weight outside the assumed limits (e.g. analysis of PC5).

The totals listed in parentheses are dry weight totals where the true volatile contents are unknown and high totals are almost certainly due to sample and flux weighing errors and bead inhomogeneity.

Ten trace elements out of the eleven listed in Table II.2 were analysed by XRF spectrometry at Oxford on pressed powder pellets made up from the same samples as those listed in Table II.1. Chromium was analysed for by flame atomic absorption in Oxford by Miss P. Shreeve.

The elements Rb, Sr, X and Zr were analysed together in the same run with the same machine settings, so that inter-element interferences and background effects could be taken into account using the correction procedures due to Pankhurst (1969) and a computer program 'CA5'.

The mass adsorption coefficients derived from the measurement of the Compton scatter peak in the Rb, Sr, Y, Zr run were also used for the correction of Nb raw count data, but for this element the major source of uncertainty is standard concentration data. The formula used to derive Nb concentration was:

$$\text{Nb p.p.m.} = 5.621 \times (\text{MAC at MoK}\alpha) \times x + 3.8$$

(assuming USGS standard GSP-1 has 29 p.p.m. Nb) where x is the count rate ratio of unknown to standard.

Those Nb analyses in parentheses were not corrected for mass adsorption effects, but the raw counts were calibrated against a range of USGS standards and an Icelandic alkali basalt I63/31 supplied by I. Friedleiffson (Nb = 47 p.p.m.).

Ce, Nd and Ba were analysed together on the same run, and no attempt was made to correct for mass adsorption or inter-element interference since

the USGS standards W-1, BCR-1, AGV-1 and G-2 provided a reasonable uncorrected linear calibration in the range of the analysed samples.

La and Ni were also analysed without recourse to corrections, and Ni contents were calculated using the formula:

$$\text{Ni p.p.m.} = 88.2 x + 12.6$$

(assuming USGS standard W-1 has 88.2 p.p.m. Ni) where x is the count rate ratio of unknown to standard.

For the elements Rb, Sr, Y and Zr precision is estimated from duplicate analyses to be of the order of $\pm 3\%$, except for Rb analyses of less than 20 p.p.m. which are only precise to within $\pm 10\%$. Accuracy has been estimated from the analysis of USGS standard rocks:

p.p.m. BCR-1 standard value		p.p.m. W-1 This study	p.p.m. W-1 recommended value
Rb	47.3	20.8 $\pm$ 1	21
Sr	332.1	196.0 $\pm$ 10	190
Y	37.1	21.8 $\pm$ 3	25
Zr	190.	100 $\pm$ 5	105

At concentrations greater than 20 p.p.m., the values obtained for W-1 in this study fall within 5% deviation of the recommended values (Flanagan, 1973). La, Ce, Nd and Ba are only accurate to ca. $\pm 10\%$, though Ce/Nd ratios are probably both more precise and accurate ($\pm 5\%$) judging by comparisons of this XRF method with isotope dilution methods (S. Carter, pers. comm., 1976).

The precision of the Ni analyses is estimated at $\pm 5\%$, though great uncertainties remain about the accuracy due to the lack of mass adsorption correction.

In all cases, apart from Rb, detection limits are around 5 p.p.m. Standard concentration values used are those given in Pankhurst and O'Nions (1973), Flanagan (1973) and Nakamura (1974).

Artificial standards were used for chromium analysis, and the precision is estimated at $\pm 3\%$, increasing to $\pm 15\%$ for very low concentrations (i.e. 10 p.p.m.).

Table II.1. XRF fusion bead analyses of major elements in rocks from the Ile de Groix. % weight oxides

	LM12	PC10	PC7	TUD5	PC5	TRE1	TRE2	PM10	KB1	PN1
SiO ₂	49.38	47.02	48.04	47.37	53.21	49.50	49.47	52.62	51.72	45.51
Al ₂ O ₃	14.36	13.99	14.49	16.23	15.18	15.45	17.83	17.61	14.88	15.46
TiO ₂	1.48	2.26	2.75	2.21	1.83	1.18	1.57	2.31	2.24	1.76
FeO	7.72	11.84	9.57	5.71	7.89	7.25	6.55	7.12	5.03	3.82
Fe ₂ O ₃	4.44	4.52	5.12	8.58	5.53	5.39	6.86	6.21	8.14	10.10
MnO	0.20	0.23	0.20	0.26	0.19	0.18	0.27	0.19	0.12	0.12
CaO	9.64	10.63	9.08	11.59	6.18	8.73	8.98	7.24	6.85	11.57
MgO	7.44	4.95	6.00	4.12	6.16	5.15	5.84	3.04	2.60	3.58
K ₂ O	0.03	0.34	0.25	0.16	0.06	0.04	0.54	0.22	0.44	1.84
Na ₂ O	4.48	3.52	4.50	2.60	4.71	4.17	3.34	2.24	5.13	3.71
P ₂ O ₅	0.11	0.33	0.33	0.34	0.22	0.19	0.06		0.86	0.34
'H ₂ O'	0.78	0.63		0.96	1.17				1.09	1.12
CO ₂ [†]		0.04		0.29						3.92
TOTAL	100.04	100.30	(100.33)	100.42	102.33*	(97.24)	(101.31)	(98.77)	99.12	102.85
Loss on fusion	1.32	0.35	1.65	2.29	1.55	1.79	1.66	2.41	1.63	6.08

*Flux weight error † determined by M. Hepher, gravimetric method Error ± 5%

Table II.1. XRF fusion bead analyses of major elements in rocks from the Ile de Groix. % weight oxides (contd.)

	GAD1	GAD2	PC1	LIN-1	KERZ 1	BIL3	BIL10	BIL7	TUD6	TUD7
SiO ₂	52.22	48.60	47.55	48.53	48.59	56.11	48.66	50.53	71.78	51.08
Al ₂ O ₃	15.86	15.16	15.11	16.38	16.43	12.43	15.89	16.30	13.97	15.54
TiO ₂	1.51	1.79	1.12	1.76	2.30	0.56	2.45	1.65	0.26	2.32
FeO	3.21	8.16	6.54	7.10	5.09	3.35	5.72	5.70		5.81
Fe ₂ O ₃	8.25	3.33	2.33	2.54	4.62	2.68	8.74	5.27	2.51 [‡]	6.60
MnO	0.21	0.15	0.14	0.18	0.13	0.13	0.17	0.22	0.16	0.10
CaO	13.33	6.89	9.75	6.93	9.90	6.98	9.01	4.69	1.27	0.54
MgO	2.76	9.28	12.69	8.25	7.92	9.44	4.39	4.83	0.69	7.24
K ₂ O	0.22	0.10	0.17	0.09	1.20	0.37	0.54	2.61	3.27	1.19
Na ₂ O	2.25	3.70	1.97	4.75	2.48	5.67	3.10	2.03	3.50	1.99
P ₂ O ₅	0.26	0.19	0.12	0.33	0.46	0.35	0.20	0.67	0.15	0.42
'H ₂ O'				2.76		1.22		3.28		6.65
CO ₂				0.56						
TOTAL	(100.08)	(97.36)	(97.50)	100.16	(99.12)	99.29	(98.87)	97.73	(97.56)	99.48
Loss on fusion	1.63	3.46	3.2	4.15	3.43	2.13	5.07	6.58	2.67	7.20

[‡]Total Fe as Fe₂O₃, analysis by P. Webb

Table II.2. Trace element analyses of rocks from the Ile de Groix. Contents in p.p.m.

Sample No.→	LM12	PC10	PC7	TUD5	PC5	TRE1	TRE2	PM10	KB1	PN1
Rb	0.1	8.6	6.8	4.3	3.3	0.8	16.0	5.9	13.1	36.8
Sr	116.7	111.7	109.9	261.6	161.8	208.2	369.8	302.4	239.7	253.3
Y	36.7	66.4	62.4	56.5	45.7	43.4	52.6	48.4	40.9	28.6
Zr	88.2	219.6	227.8	215.5	145.7	140.6	114.0	155.4	216.4	158.0
Nb	n.d.	9.0	(5)	7.2	(5)	(5)	n.d.	(6.5)	36.0	23.0
Ce	17.0	34.0	31.8	21.5	27.0	23.0	23.9	28.0	51.8	42.8
Nd	11.7	24.0	21.6	19.3	18.5	15.9	16.2	18.9	31.1	21.9
Ba	21	65.2	65.6	35.0	46.8	24.0	103.0	35.0	114.3	168.8
La	n.d.	7	9	n.d.	9	9	n.d.	8	20	16
Ni	60.3	74.5	78.6	153.0	28.0	37.1	28.2	140.4	5	16.6
Cr	219	142.5	120	262.5	30	76	39	248	12	21

n.d. not detectable

Table II.2. Trace element analyses of rocks from the Ile de Groix. Contents in p.p.m.
(contd.)

Sample No→	GAD1	GAD2	PC1	LIN-1	KERZ 1	BIL3	BIL10	BIL7	TUD6	TUD7
Rb	3.0	0.3	5.1	0.9	24.4	12.6	21.3	82.4	149.6	40.8
Sr	372.9	92.1	120.6	310.8	477.3	361.8	181.7	104.6	182.7	21.5
Y	34.5	33.2	23.3	33.7	28.0	20.8	45.3	51.9	-	30.3
Zr	108.6	130.7	88.5	183.2	197.3	124.3	145.2	288.8	-	243.6
Nb	(5.4)	(6.5)	(6.0)	24.7	(35.0)	10.4	(4.2)	17.5	-	22.4
Ce	61.7	24.0	11.4	40.8	49.3	31.9	27.0	74.0	-	41.1
Nd	29.7	13.3	9.0	22.0	25.0	16.2	15.6	37.8	-	21.6
Ba	131.8	26.0	8.8	50.7	260.0	109.8	135	412.0	-	70.3
La	5	n.d.	n.d.	13	20	13	5	30	-	14
Ni	93.7	130.0	422.0	123.5	182.3	221.7	97.2	21.1	-	212.0
Cr	208.0	226.0	701.0	189.0	276.0	191.0	218.0	38.0	-	-

n.d. not detectable

Table II.3. Machine settings used for XRF analysis of ten trace elements

ELEMENT	LINE	2θ° peak	TUBE	KV/mA	Crystal	Counter	EHT _v	Coll.	LL	W	2θ° Backgrounds	
Rb	Kα	37.99	Mo	100/20	LiF 220	Scint.	950	Coarse	80	120	37.04	38.94
Sr	Kα	35.85	Mo	"	"	"	"	"	"	"	34.66	
Y	Kα	33.88	Mo	"	"	"	"	"	"	"	33.10	
Zr	Kα	32.00	Mo	"	"	"	"	"	"	"	Compton scatter pK. 30.04 Bg min = 31.61	
Nb	Kα	30.42	W	100/20	LiF 220	Scint.	900	Coarse	60	60	29.88	33.08
Ce	Lβ ₁	71.62	W	95/24	LiF 200	Flow	1500	Fine	40	40	70.80	74.30
Nd	Lα	72.11	W	"	LiF 200	"	"	"	"	"	"	"
Ba	Lβ ₂	73.32	W	"	LiF 200	"	"	"	"	"	"	"
La	Lα	138.99	W	"	LiF 220	"	1650	Coarse	200	100	135.99	141.99
Ni	Kα	71.18	W	75/24	LiF 220	Scint.	1025	Fine	200	150	69.18	73.18

100 sec.
counting
times

40 sec.
counting times

Fig. II.1

, An AFM plot (Wt. % oxides) of the basic schists from the Ile de Groix.

- Zone I
- Zone II
- Albitic micaschist

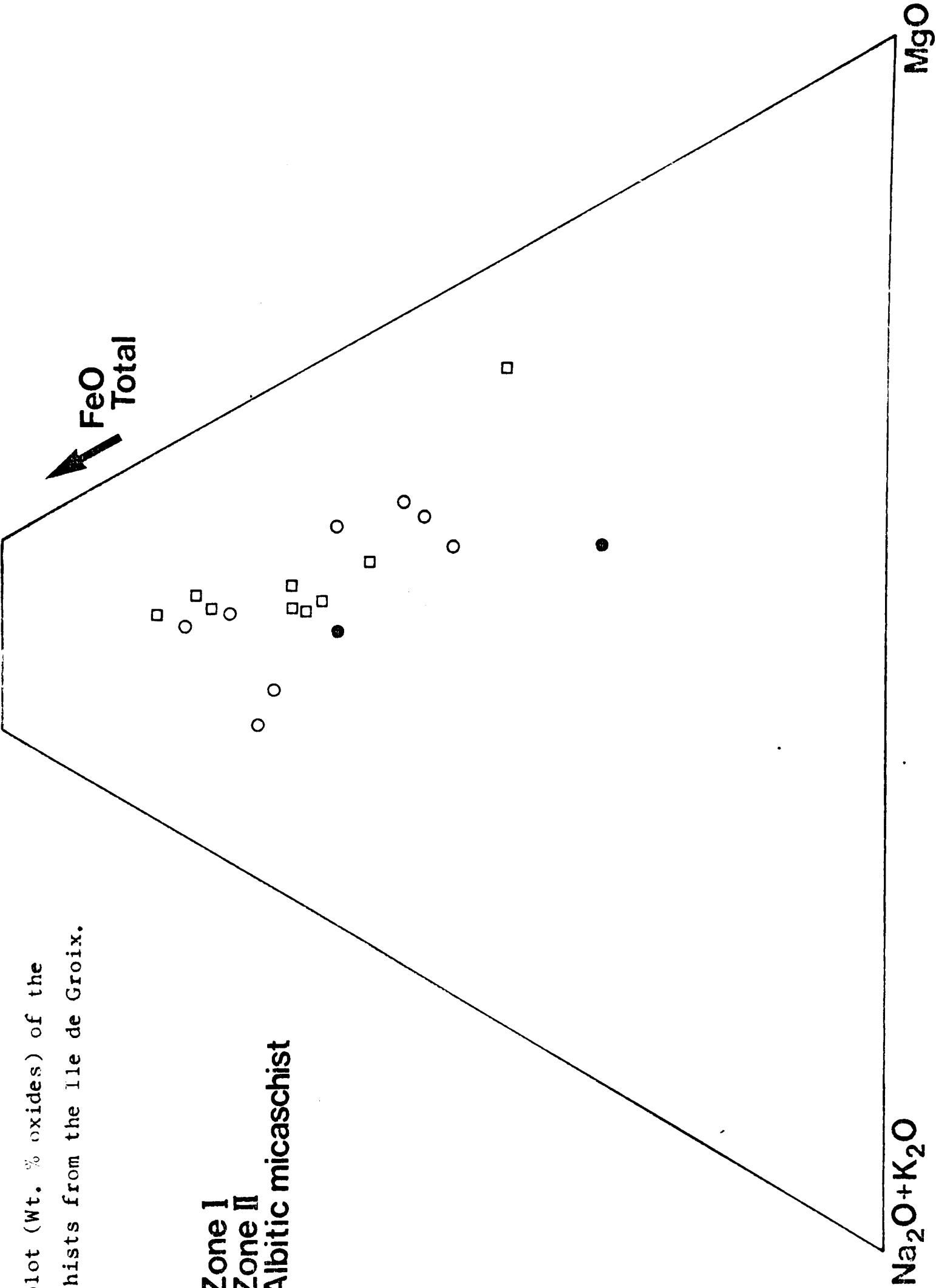


Fig. II.2

Histogram of TiO_2 content in metabasics from the Ile de Groix. (Data of Makanjuola 1967, Carpenter 1976).

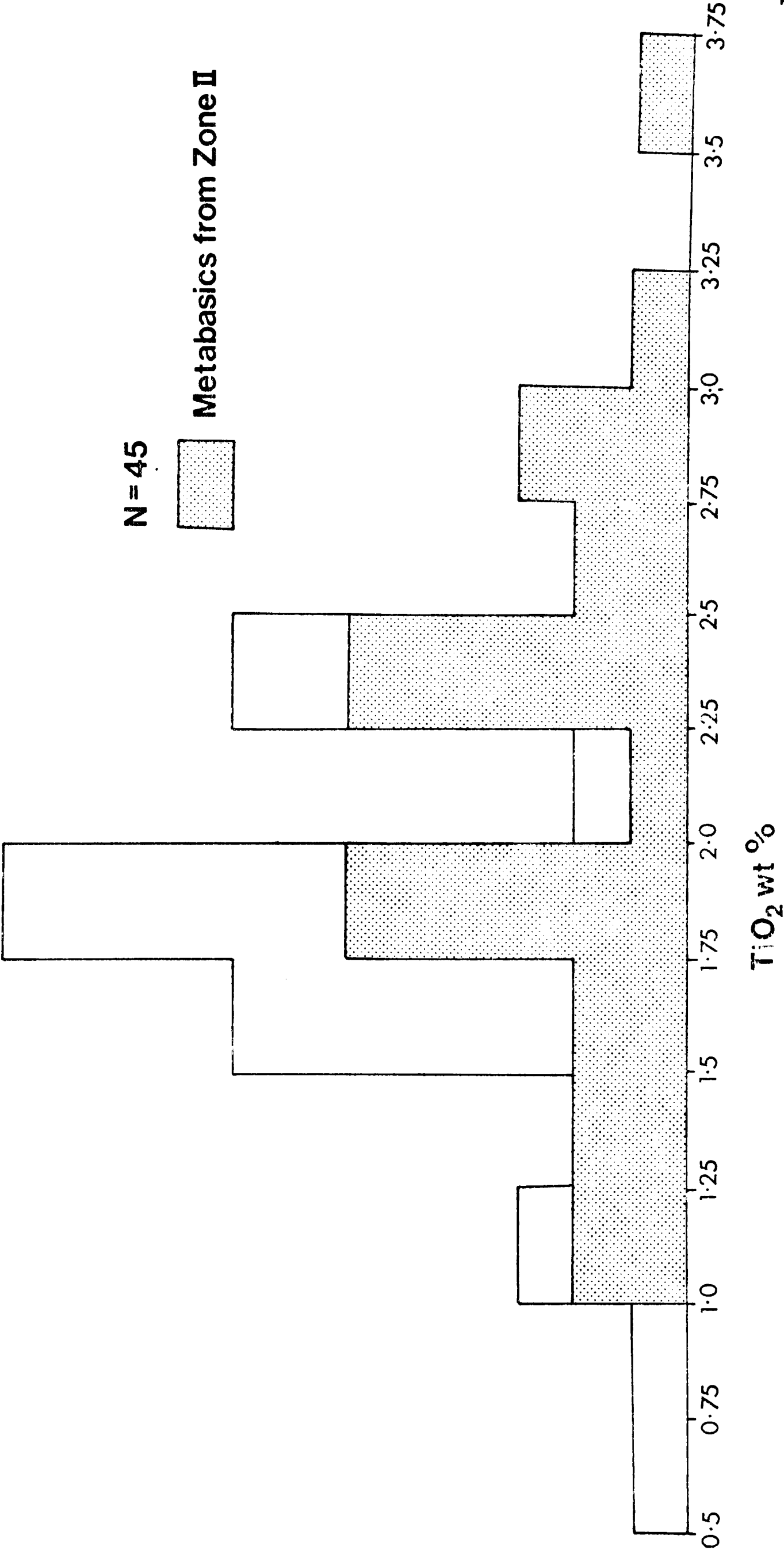


Fig. 11.3

Ti - Zr - Y discriminant diagram
(after Pearce and Cann, 1973) for
Groix metabasics.

- Zone I
- Zone II

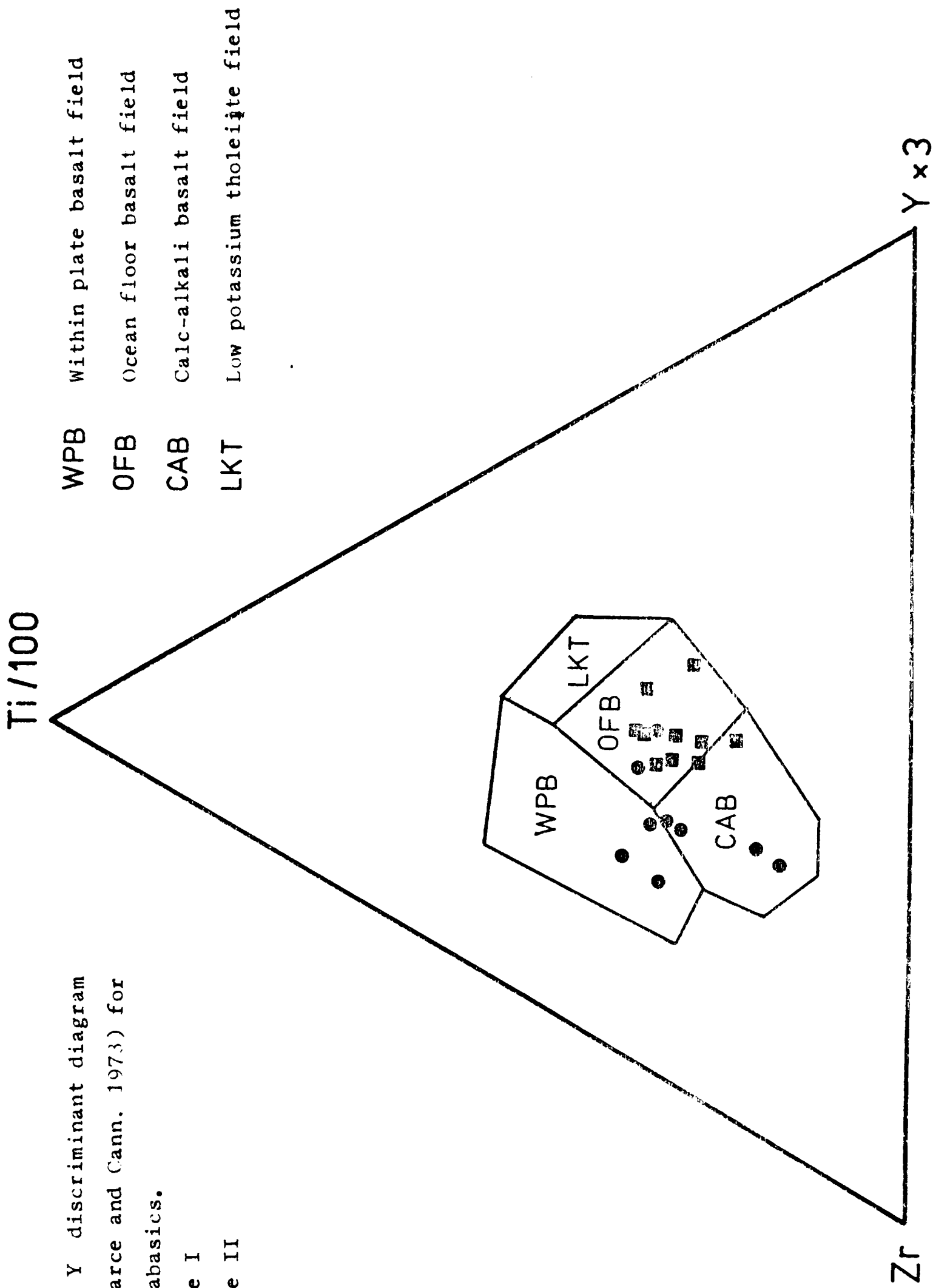


Fig. II.4

REE patterns from zone II Groix metabasites. Abundances normalised to chondrite values of Nakamura (1974).

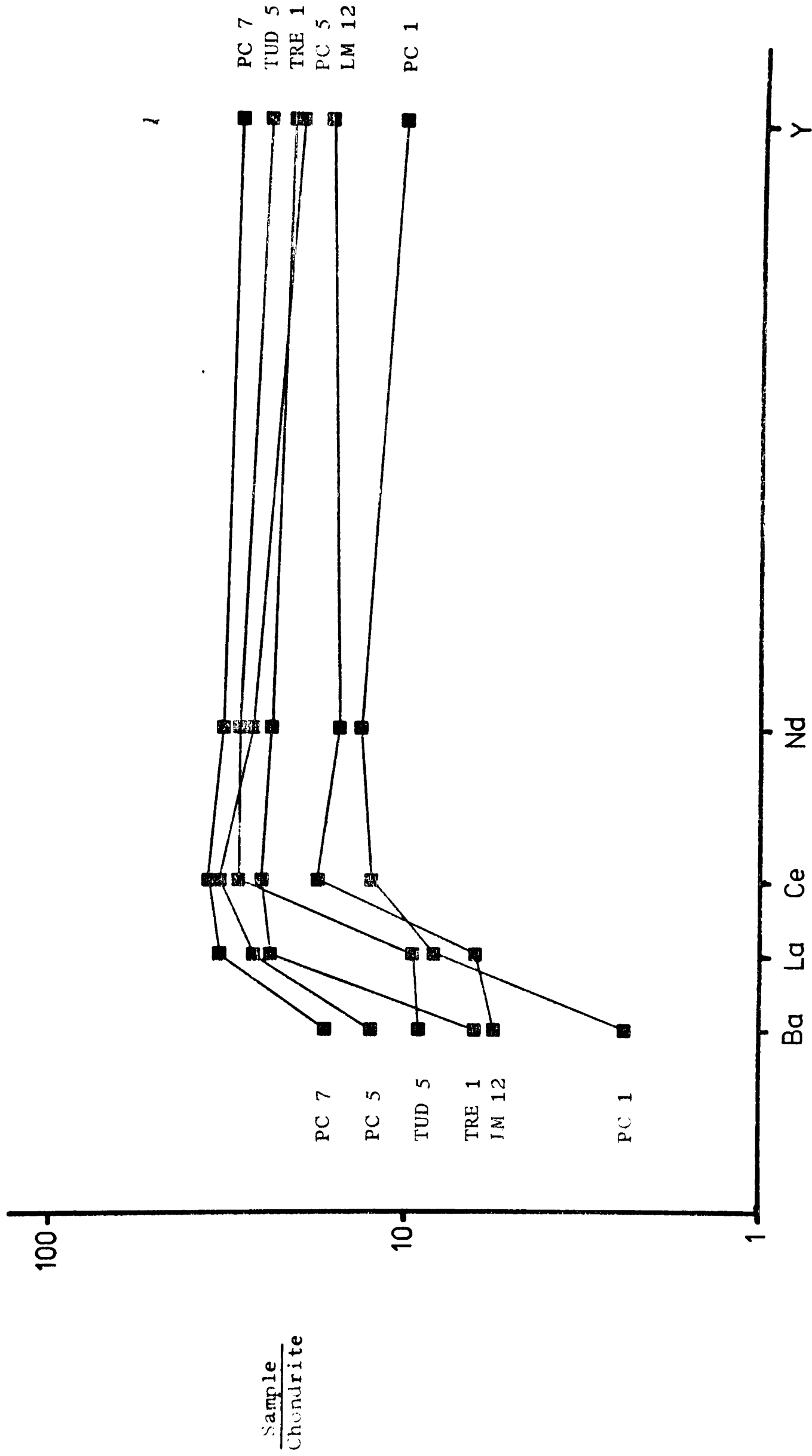
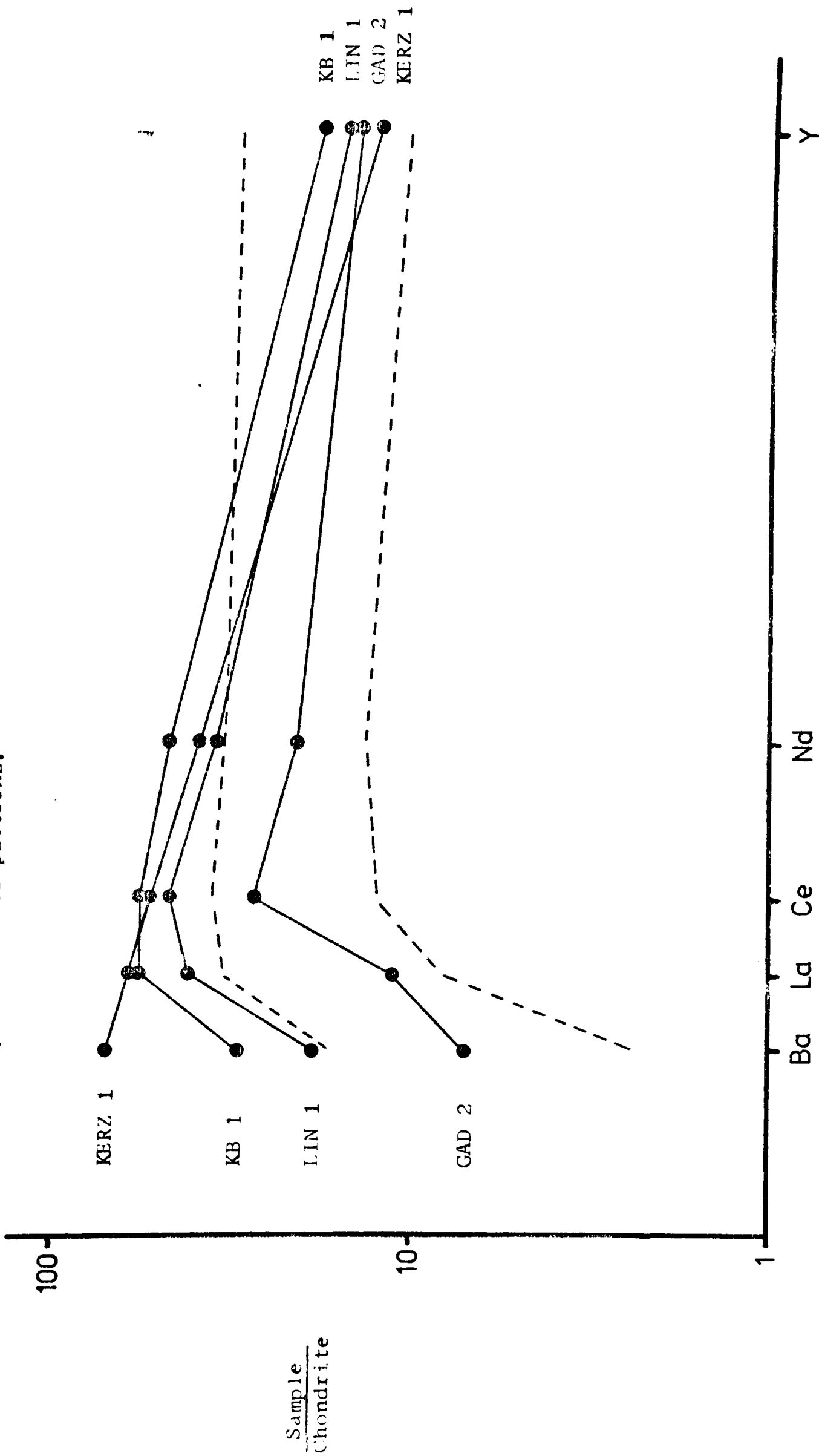


Fig. II.5

Rb patterns from zone I Groix metabasites. Abundances normalised to chondrite values of Nakamura (1974).
 Dashed curves show envelope of zone II patterns.



APPENDIX III

Electron Probe Analyses of Minerals from the Ile de Groix

The analyses quoted in this appendix were performed on three separate instruments:-

1. The Geoscan MkII electron microprobe at the Department of Mineralogy and Petrology, Cambridge between May and September 1974, under the supervision of Dr. J.V.P. Long and Mr. G.G. Price.
2. The Harwell Energy-Dispersive Spectrometer microprobe at the Department of Mineralogy and Petrology, Cambridge, under the supervision of Drs. J.V.P. Long and P. Statham, during March 1975. Analyses performed on this instrument are suffixed "E.D.S.".
3. The Microscan 9 Cambridge Instruments electron microprobe at the Department of Geology and Mineralogy, Oxford, under the supervision of Dr. W.A.P. Nicholson, during January 1976. Analyses performed on this instrument are suffixed "MS9".

Specimen preparation and analysis correction (ZAF) techniques are those described in Sweatman and Long (1969). For the E.D.S. probe, spectrum processing was effected by the program PESTSTRIP 2, and the ZAF corrections are similar to those in the program TIM3, with mass absorption coefficients fitted to Heinrich's (1966) data.

Standards used, and other relevant run data for the Geoscan probe, using a constant accelerating voltage of 20 KV, were:-

	Standard	% wt element	10 ⁻⁹ amp, Probe current Peak rate cps.		Spectrometer crystal
Na	Jadeite	11.34	30	260*	KAP
Al	Corundum	52.93	40	4500†	KAP
K	Orthoclase	12.40	30	3500*	PET
Ca/Si	Wollastonite	Ca 34.32	30	9800*	PET
		Si 24.00	40	2600†	KAP
Ti	Pure Ti metal	100	40	6800	PET
Mn	Pure Mn metal	100	40	16800	LiF
Fe	Pure Fe metal	100	40	17000	LiF
Mg	Olivine (St. John's Is)	29.67	30	1500†	KAP

* Mylar window out. † Gas flow proportional counter, other elements analysed with sealed (Xe) counter.

Apparent element concentrations were calculated from the raw count data assuming a dead-time of 2×10^{-6} sec, and then were corrected for absorption and fluorescence interference using Heinrich's (1966) set of mass absorption coefficients.

The precision of the analyses varies with both the atomic number and the concentration of the element in question, and has been estimated by D.J. Waters (1976) to be of the order $\pm 3 - 10\%$ (2σ error) for a range of Na apparent concentration 10 - 1 wt% and $\pm 0.5 - 4\%$ (2σ) for a range of Fe apparent concentration 10 - 1 wt% using 10 sec counting times. Generally, precision calculated from the variation in standard count rates is better than $\pm 5\%$ for most major elements occurring in concentrations greater than 10 apparent wt%.

A useful check on the accuracy of the analyses is provided by the analysis of the St. John's olivine standard for Fe and Si.

Apparent ZAF uncorrected element wt%

	This study	Recommended value
Fe	6.48	6.41 ± 0.2
Si	15.31_6	15.59 ± 0.4

With the E.D.S. probe, detection limits are generally higher (especially for the light elements Na, Mg, Al, Si) but precision and accuracy are comparable with most conventional crystal dispersive microprobes, except that Si is probably over-estimated by 0.5% due to a discontinuity in the background level at this wavelength (P. Statham, pers. comm. 1975).

Pure cobalt metal was used as the primary standard in the PESTSTRIP 2 program for calculating peak areas of the analysed elements, and secondary standards were run as a check on accuracy:

ZAF Corrected		element wt%	
		This study (E.D.S.)	Recommended value
St. John's olivine	(Mg	29.41	29.67
	(Fe	7.590	7.415
	(Si	18.85	19.07
Wollastonite	(Ca	34.21	34.32
	(Si	23.51	24.00
Orthoclase	K	12.01	12.40
Corundum	Al	51.54	52.93

A probe current of approx. 15×10^{-9} Amp. was used throughout E.D.S. analysis, with an accelerating voltage of 20 KV.

The standards used for the MS9 probe runs were as follows, with element concentrations derived from stoichiometric proportions unless otherwise stated.

Element	Standard	% wt element	Spectrometer crystal		Counter EHT in KV
Na	Jadeite	10.96 [†]	RbAP	} Flow counter with 1 μ window	1.525
Mg	Periclase	60.31	RbAP		
Al	Corundum	52.93	RbAP		
Si	Quartz	46.75	RbAP		
K	Orthoclase	13.44	PET	} Flow counter with 6 μ Mylar window	1.450
Ca	Wollastonite	34.20	PET		
Ti	Rutile	59.95	PET		
Mn	Rhodonite	32.18	LiF ₂₀₀	} Flow counter with 6 μ Mylar window	1.450
Fe	Haematite	69.94	LiF ₂₀₀		

[†] Average of 28 wet chemical analyses, P. Shreeve.

An accelerating voltage of 20 KV, and a probe current of 20×10^{-9} Amp. were used throughout.

The analyses were not normalized to total 100% (dry basis) and were recalculated into cation proportions and end-member molecular percentages

using the computer program RECALC, modified by D.J. Waters (1976) for the purpose of yielding pyroxene end-members consistent with the scheme of Powell and Powell (1974).

Note: Fe has been calculated for glaucophanes, micas and chlorites using an arbitrary, fixed fraction of Fe_{TOTAL} as Fe_2O_3 , chosen on the basis of averaged wet chemical analyses (Makanjuola and Howie, 1972; Triboulet, 1974).

	$Fe_2O_3 / Fe_2O_3 + FeO$
Glaucophanes	0.31
Micas	0.705
Chlorites	0.206

For epidotes, all Fe is calculated as Fe_2O_3 , and for garnets, Fe^{3+} (for andradite molecule) has been apportioned on a site occupancy basis ($Si + Al + Fe^{3+} = 5$).

PC10

UNBPS9

PC10

GTBPS9

PC10

GTA MS9

OXIDE WT %
SI02 53.23
AL2O3 7.15
TI02 0.04
FE2O3 8.03
FEO 2.00
MGO 2.70
MNO 0.00
CAO 0.99
NA2O 13.38
K2O 0.03

TOTAL 98.21

RECALCULATED ON
BASIS OF 100.0 (0)

SI 1.975
AL 0.025
AL 0.287
TI 0.001
FE 0.168
FE 0.112
FE 0.129
MGO 0.559
MNO 0.459
CAO 0.001

ENDMEMBERS

% DIOPS 42.90
% HED 11.17
% CATSCH 10.00
% JAD 31.39
% AC 16.83
% TI 0.11
% H/H+F 79.34

OXIDE WT %
SI02 37.43
AL2O3 21.60
TI02 0.03
FE2O3 8.03
FEO 2.70
MGO 2.70
MNO 0.00
CAO 0.99
NA2O 10.00
K2O 0.02

TOTAL 101.54

RECALCULATED ON
BASIS OF 100.0 (0)

SI 2.941
AL 0.059
AL 1.942
TI 0.001
FE 0.055
FE 0.039
FE 0.056
MGO 0.255
MNO 0.261
CAO 0.861

ENDMEMBERS

% PYROPE 8.29
% ALM 61.09
% SPSS 0.80
% ANDR 0.74
% GROSS 25.11
% H/H+F 11.05

OXIDE WT %
SI02 36.01
AL2O3 21.65
TI02 0.11
FE2O3 8.03
FEO 2.70
MGO 2.70
MNO 0.00
CAO 0.99
NA2O 0.00
K2O 0.01

TOTAL 100.61

RECALCULATED ON
BASIS OF 100.0 (0)

SI 2.930
AL 0.070
AL 1.956
TI 0.007
FE 0.035
FE 0.034
FE 0.034
MGO 0.234
MNO 0.219
CAO 0.846

ENDMEMBERS

% PYROPE 7.54
% ALM 62.27
% SPSS 0.61
% ANDR 3.12
% GROSS 25.12
% H/H+F 10.80

PC10 011A 1159

OXIDE	WT %
SI02	53.52
AL2O3	38.15
TiO2	0.04
FE0	5.87
MGO	4.05
MNO	6.84
CAO	0.03
NA2O	12.89
K2O	16.97
	0.03
TOTAL	98.30

RECALCULATED ON
BASIS OF 6.0 (0)

SI	1.979
AL	0.021
AL	0.334
TI	0.001
FE3+	0.163
FE2+	0.125
MGO	0.377
MNO	0.001
CA	0.509
NA	0.500
K	0.001

ENDMEMBERS

% DIOPS	37.68
% HFD	12.52
% CATSCH	16.00
% JD	35.68
% AC	16.33
% TI=PX	10.11
% H/H+F	75.00

PC10 011B MS9

OXIDE	WT %
SI02	53.62
AL2O3	38.64
TiO2	0.06
FE0	5.09
MGO	3.76
MNO	6.71
CAO	0.00
NA2O	12.42
K2O	17.27
	0.02
TOTAL	98.49

RECALCULATED ON
BASIS OF 6.0 (0)

SI	1.975
AL	0.025
AL	0.350
TI	0.002
FE3+	0.166
FE2+	0.116
MGO	0.368
MNO	0.001
CA	0.490
NA	0.519
K	0.001

ENDMEMBERS

% DIOPS	36.84
% HFD	11.58
% CATSCH	10.00
% JD	37.59
% AC	16.61
% TI=PX	10.17
% H/H+F	76.08

PC10 01C MS9

OXIDE	WT %
SI02	52.95
AL2O3	38.71
TiO2	0.02
FE0	8.00
MGO	1.91
MNO	8.07
CAO	0.34
NA2O	14.63
K2O	0.02
TOTAL	98.76

RECALCULATED ON
BASIS OF 6.0 (0)

SI	1.957
AL	0.043
AL	0.249
TI	0.001
FE3+	0.223
FE2+	0.059
MGO	0.446
MNO	0.002
CA	0.548
NA	0.475
K	0.001

ENDMEMBERS

% DIOPS	44.55
% HFD	5.90
% CATSCH	0.00
% JD	29.55
% AC	22.30
% TI=PX	0.06
% H/H+F	88.31

PC10 ONE HS9

OXIDE	WT %
SI02	53.88
AL2O3	38.63
TiO2	0.04
FE2O3	0.79
FE0	3.35
MGO	7.20
MNO	0.00
CAO	12.87
NA2O	2.14
K2O	0.02
TOTAL	98.92

RECALCULATED ON
BASIS OF 6.0 (O)

SI	1.972
AL	0.028
AL	0.345
TI	0.001
FE3+	0.160
FE2+	0.103
FE0	0.390
MGO	0.502
MNO	0.502
CA	0.001
NA	
K	

ENDMEMBERS

% DIOPS	39.61
% HED	10.20
% CATSCH	0.00
% JD	32.41
% AC	15.05
% TI=PX	0.11
% H/H+F	79.43

PC10 ONE HS9

OXIDE	WT %
SI02	53.63
AL2O3	37.03
TiO2	0.23
FE2O3	6.00
FE0	2.50
MGO	7.50
MNO	0.04
CAO	13.00
NA2O	2.02
K2O	0.00
TOTAL	98.85

RECALCULATED ON
BASIS OF 6.0 (O)

SI	1.970
AL	0.030
AL	0.313
TI	0.025
FE3+	0.179
FE2+	0.111
FE0	0.401
MGO	0.531
MNO	0.531
CA	0.001
NA	0.001
K	

ENDMEMBERS

% DIOPS	41.05
% HED	18.00
% CATSCH	0.00
% JD	34.30
% AC	17.50
% TI=PX	0.17
% H/H+F	82.18

LM12 GT1 EDS

OXIDE	WT %
SI02	36.70
AL2O3	20.05
TiO2	0.18
FE2O3	0.77
FE0	2.59
MGO	1.80
MNO	0.22
CAO	10.63
NA2O	0.00
K2O	0.00
TOTAL	97.04

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.005
AL	
AL	1.935
TI	0.011
FE3+	0.047
FE2+	0.220
FE0	0.222
MGO	0.032
MNO	
CA	
NA	
K	

ENDMEMBERS

% PYROPE	7.32
% ALM	60.57
% SPSS	0.75
% ANDRS	27.08
% GROSS	10.78
% H/H+F	

PG10 GTA HS9

OXIDE	WT %
SiO2	36.91
Al2O3	21.65
TiO2	0.11
Fe2O3	0.59
FeO	0.13
MgO	1.98
MnO	0.28
CaO	0.95
Na2O	0.00
K2O	0.01
TOTAL	100.61

RECALCULATED ON
BASIS OF 12.0 (O)

SI	2.930
AL	0.070
AL	1.050
TI	0.007
FE3+	0.835
FE2+	1.234
MG	0.219
MN	0.840
CA	0.001
NA	
K	

ENDMEMBERS

% PYROPE	7.54
% ALH	0.27
% SPHSS	0.61
% GROSS	3.12
% M/H+F	10.80

TUD5 GTD MS9

OXIDE	WT %
SiO2	37.44
Al2O3	21.16
TiO2	0.00
Fe2O3	0.00
FeO	28.89
MgO	1.41
MnO	1.82
CaO	8.82
Na2O	0.04
K2O	0.01
TOTAL	99.61

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.003
AL	
AL	2.001
TI	0.001
FE3+	
FE2+	1.938
MG	0.149
MN	0.124
CA	0.758
NA	0.001
K	

ENDMEMBERS

% PYROPE	5.62
% ALH	0.59
% SPHSS	4.12
% GROSS	0.08
% M/H+F	28.00

TUD5 GTZ HS9

OXIDE	WT %
SiO2	37.61
Al2O3	20.77
TiO2	0.03
Fe2O3	0.23
FeO	29.31
MgO	1.58
MnO	1.78
CaO	8.28
Na2O	0.00
K2O	0.00
TOTAL	99.61

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.018
AL	
AL	1.965
TI	0.002
FE3+	0.014
FE2+	1.967
MG	0.189
MN	0.121
CA	0.712
NA	0.003
K	

ENDMEMBERS

% PYROPE	6.26
% ALH	6.18
% SPHSS	4.01
% GROSS	0.23
% M/H+F	28.76

TUD5 GTV MS9

OXIDE	WT %
SI02	37.33
AL2O3	20.94
TiO2	0.07
FE0	0.50
MGO	9.85
MNO	1.49
CAO	1.60
NA2O	8.40
K2O	0.00
TOTAL	100.18

RECALCULATED ON
BASIS OF 12.0 (O)

SI	2.988
AL	0.012
AL	1.065
Ti	0.004
FE3+	0.030
FE2+	1.009
MG	0.178
MN	0.108
CA	0.121
NA	
K	

ENDMEMBERS

% PYROPE	5.89
% ALI	6.25
% SPSS	3.60
% ANDR	1.70
% GROSS	22.12
% H/H+F	28.17

TUD5 GTW MS9

OXIDE	WT %
SI02	37.37
AL2O3	20.74
TiO2	0.07
FE0	0.86
MGO	9.79
MNO	1.40
CAO	1.33
NA2O	8.40
K2O	0.00
TOTAL	100.53

RECALCULATED ON
BASIS OF 12.0 (O)

SI	2.987
AL	0.013
AL	1.942
Ti	0.002
FE3+	0.052
FE2+	1.091
MG	1.167
MN	0.125
CA	
NA	
K	

ENDMEMBERS

% PYROPE	5.52
% ALI	6.03
% SPSS	4.10
% ANDR	2.86
% GROSS	21.16
% H/H+F	27.73

TUD5 GTA MS9

OXIDE	WT %
SI02	39.06
AL2O3	20.75
TiO2	0.06
FE0	0.00
MGO	26.85
MNO	1.40
CAO	1.78
NA2O	8.34
K2O	0.00
TOTAL	99.23

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.110
AL	
AL	1.948
Ti	0.002
FE3+	
FE2+	1.788
MG	0.166
MN	0.187
CA	0.712
NA	
K	0.001

ENDMEMBERS

% PYROPE	5.54
% ALI	5.60
% SPSS	6.25
% ANDR	0.46
% GROSS	23.86
% H/H+F	28.50

PC10 OMPH1			PC10 OMPH2			PC10 OMPH3		
OXIDE		WT %	OXIDE		WT %	OXIDE		WT %
SI02		53.77	SI02		54.40	SI02		54.06
AL2O3		7.86	AL2O3		7.79	AL2O3		6.40
TI02		0.12	TI02		0.00	TI02		0.00
FE2O3		7.67	FE2O3		6.00	FE2O3		6.58
FE0		3.50	FE0		3.59	FE0		3.02
MGO		5.44	MGO		6.88	MGO		7.66
MNO		0.00	MNO		0.00	MNO		0.00
CAO		11.36	CAO		12.76	CAO		13.70
NA2O		7.84	NA2O		7.09	NA2O		6.54
K2O		0.00	K2O		0.00	K2O		0.00
TOTAL		97.56	TOTAL		98.52	TOTAL		97.95

RECALCULATED ON BASIS OF 6.0 (0)			RECALCULATED ON BASIS OF 6.0 (0)			RECALCULATED ON BASIS OF 6.0 (0)		
SI		2.003	SI		2.000	SI		2.004
AL			AL			AL		
AL		0.345	AL		0.338	AL		0.280
TI		0.003	TI			TI		
FE3+		0.215	FE3+		0.166	FE3+		0.183
FE2+		0.109	FE2+		0.110	FE2+		0.094
MG		0.302	MG		0.377	MG		0.423
MN			MN			MN		
CA		0.453	CA		0.503	CA		0.545
NA		0.567	NA		0.506	NA		0.470
K			K			K		

ENDMEMBERS			ENDMEMBERS			ENDMEMBERS		
% DIOPS		30.21	% DIOPS		37.71	% DIOPS		42.31
% HED		10.91	% HED		11.03	% HED		9.36
% CATSCH		0.00	% CATSCH		0.00	% CATSCH		0.00
% JD		34.30	% JD		34.16	% JD		28.32
% AC		21.49	% AC		16.61	% AC		18.35
% TI-PX		0.35	% TI-PX		0.00	% TI-PX		0.00
% M/M+F		73.47	% M/M+F		77.37	% M/M+F		61.89

PC10 OMPH4

OXIDE	WT %
SI02	54.17
AL2O3	8.17
TI02	0.30
FE2O3	5.61
FE0	3.47
MGO	6.65
MNO	0.30
CAO	12.22
NA2O	7.24
K2O	0.00
TOTAL	97.53

RECALCULATED ON
BASIS OF 6.0 (0)

SI	2.006
AL	
AL	0.357
TI	
FE3+	0.156
FE2+	0.107
MG	0.367
MN	
CA	0.485
NA	0.523
K	

ENDMEMBERS

% DIOPS	36.71
% HED	10.74
% CATSCH	0.43
% JD	36.32
% AC	15.64
% TI-PX	0.33
% M/H+F	77.37

PC10 OMPH5

OXIDE	WT %
SI02	54.30
AL2O3	8.41
TI02	0.13
FE2O3	5.13
FE0	4.23
MGO	6.56
MNO	0.00
CAO	12.04
NA2O	7.21
K2O	0.00
TOTAL	98.01

RECALCULATED ON
BASIS OF 6.0 (0)

SI	2.003
AL	
AL	0.366
TI	0.004
FE3+	0.142
FE2+	0.131
MG	0.361
MN	
CA	0.476
NA	0.516
K	

ENDMEMBERS

% DIOPS	34.67
% HED	12.55
% CATSCH	0.00
% JD	36.16
% AC	14.25
% TI-PX	0.37
% M/H+F	73.42

PC10 OMPH6

OXIDE	WT %
SI02	54.45
AL2O3	7.27
TI02	0.00
FE2O3	5.78
FE0	3.73
MGO	7.01
MNO	0.00
CAO	12.81
NA2O	6.88
K2O	0.00
TOTAL	97.91

RECALCULATED ON
BASIS OF 6.0 (0)

SI	2.014
AL	
AL	0.317
TI	
FE3+	0.161
FE2+	0.115
MG	0.386
MN	
CA	0.507
NA	0.493
K	

ENDMEMBERS

% DIOPS	38.61
% HED	11.54
% CATSCH	0.00
% JD	32.04
% AC	16.07
% TI-PX	0.20
% M/H+F	75.99

PC10 GLB

LM12 OMPH 1

LM12 OMPH2

OXIDE	WT %	TOTAL
SI02	53.48	96.74
AL2O3	7.87	
TI02	0.42	
FE2O3	5.01	
FE0	3.94	
MGO	6.46	
MNO	0.00	
CA0	12.63	
NA20	6.93	
K20	0.00	
TOTAL		

OXIDE	WT %	TOTAL
SI02	54.65	97.59
AL2O3	7.52	
TI02	0.00	
FE2O3	2.89	
FE0	4.13	
MGO	8.19	
MNO	0.00	
CA0	14.25	
NA20	5.95	
K20	0.00	
TOTAL		

OXIDE	WT %	TOTAL
SI02	53.57	96.17
AL2O3	7.75	
TI02	0.00	
FE2O3	2.79	
FE0	4.50	
MGO	7.62	
MNO	0.00	
CA0	13.98	
NA20	5.96	
K20	0.00	
TOTAL		

RECALCULATED ON
BASIS OF 6.0 (0)

SI	2.001
AL	
AL	0.347
TI	0.012
FE3+	0.141
FE2+	0.123
MG	0.360
MN	
CA	0.507
NA	0.503
K	

ENDMEMBERS

% DIOPS	36.03
% HED	12.31
% CATSCH	0.00
% JU	32.62
% AC	14.12
% TI-PX	1.20
% M/M+F	74.53

RECALCULATED ON
BASIS OF 6.0 (0)

SI	2.017
AL	
AL	0.327
TI	
FE3+	0.080
FE2+	0.128
MG	0.450
MN	
CA	0.564
NA	0.426
K	

ENDMEMBERS

% DIOPS	43.92
% HED	12.44
% CATSCH	0.00
% JU	32.90
% AC	0.03
% TI-PX	0.00
% M/M+F	77.92

RECALCULATED ON
BASIS OF 6.0 (0)

SI	2.011
AL	
AL	0.343
TI	
FE3+	0.079
FE2+	0.141
MG	0.426
MN	
CA	0.562
NA	0.434
K	

ENDMEMBERS

% DIOPS	42.22
% HED	13.99
% CATSCH	0.00
% JU	34.40
% AC	7.89
% TI-PX	0.00
% M/M+F	75.11

CHLORITE

TUD5 EPC MS9

OXIDE	WT %
SiO ₂	37.227
Al ₂ O ₃	23.237
TiO ₂	20.027
Fe ₂ O ₃	11.558
FeO	0.000
H ₂ O	0.044
MnO	0.344
CaO	23.443
Na ₂ O	0.003
K ₂ O	0.000
TOTAL	96.00

RECALCULATED ON
BASIS OF 12.5 (O)

Si	3.019
Al	
Al	2.219
Ti	0.204
Fe ³⁺	0.700
Fe ²⁺	
Mg	0.005
Mn	0.023
Ca ⁺	3.035
Na	0.005
K	

ENDMEMBERS

% EP	23.53
% CZ	73.96

TUD5 EPD MS9

OXIDE	WT %
SiO ₂	37.183
Al ₂ O ₃	22.353
TiO ₂	20.129
Fe ₂ O ₃	12.000
FeO	0.027
MgO	0.241
MnO	0.211
CaO	23.003
Na ₂ O	0.002
K ₂ O	
TOTAL	96.11

RECALCULATED ON
BASIS OF 12.5 (O)

Si	3.021
Al	
Al	2.141
Ti	0.008
Fe ³⁺	0.764
Fe ²⁺	
Mg	0.038
Mn	0.000
Ca	2.000
Na	0.002
K	

ENDMEMBERS

% EP	25.45
% CZ	71.36

TUD5 MCA MS9

OXIDE	WT %
SiO ₂	25.098
Al ₂ O ₃	19.433
TiO ₂	20.044
Fe ₂ O ₃	20.711
FeO	16.955
MgO	18.300
MnO	0.227
CaO	0.044
Na ₂ O	0.002
K ₂ O	0.006
TOTAL	85.80

RECALCULATED ON
BASIS OF 28.0 (O)

Si	5.450
Al	
Al	4.805
Ti	0.006
Fe ³⁺	0.743
Fe ²⁺	3.974
Mg	5.721
Mn	0.048
Ca	0.000
Na	0.008
K	0.016

ENDMEMBERS

% MGCLIN	37.21
% FECLIN	29.74
% M/M+P	65.80

PC10 GT CORE 1

OXIDE	WT %
SI02	37.29
AL2O3	20.22
TI02	0.00
FE2O3	1.12
FE0	29.48
MGO	1.32
MNO	0.73
CAO	9.32
NA2O	0.00
K2O	0.00
TOTAL	99.47 ✓

RECALCULATED ON
BASIS OF 12.0 (0)

SI	3.007
AL	
AL	1.922
TI	
FE3+	0.068
FE2+	1.988
MG	0.158
MN	0.050
CA	0.805
NA	
K	

ENDMEMBERS

% PYROPE	5.24
% ALM	65.92
% SPLESS	1.65
% ANDR	3.38
% GROSS	23.32
% M/M+F	7.37

PC10 GT RIM 1

OXIDE	WT %
SI02	37.46
AL2O3	20.05
TI02	0.00
FE2O3	0.97
FE0	28.02
MGO	2.02
MNO	0.22
CAO	9.66
NA2O	0.00
K2O	0.00
TOTAL	98.41

RECALCULATED ON
BASIS OF 12.0 (0)

SI	3.028
AL	
AL	1.911
TI	
FE3+	0.059
FE2+	1.894
MG	0.244
MN	0.015
CA	0.836
NA	
K	

ENDMEMBERS

% PYROPE	8.03
% ALM	62.39
% SPLESS	0.51
% ANDR	2.90
% GROSS	24.65
% M/M+F	11.40

PC10 GT/CHL RIM

OXIDE	WT %
SI02	37.25
AL2O3	19.79
TI02	0.00
FE2O3	1.29
FE0	27.44
MGO	2.07
MNO	0.23
CAU	9.93
NA2O	0.00
K2O	0.00
TOTAL	98.00

RECALCULATED ON
BASIS OF 12.0 (0)

SI	3.024
AL	
AL	1.894
TI	
FE3+	0.079
FE2+	1.863
MG	0.251
MN	0.016
CA	0.864
NA	
K	

ENDMEMBERS

% PYROPE	8.26
% ALM	61.39
% SPLESS	0.53
% ANDR	3.91
% GROSS	24.57
% M/M+F	11.00

PC10 GT RIM 2A

OXIDE	WT %
SI02	37.49
AL2O3	19.63
TiO2	0.00
FE2O3	1.98
FE0	27.29
MGO	2.33
MNO	0.33
CAO	10.01
NA2O	0.00
K2O	0.00
TOTAL	99.06

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.014
AL	
AL	1.861
TI	
FE3+	0.120
FE2+	1.836
MG	0.279
MN	0.023
CA	0.863
NA	
K	

ENDMEMBERS

% PYROPE	9.22
% ALM	60.59
% SPSS	0.75
% ANDR	5.93
% GROSS	22.55
% M/M+F	13.20

PC10 GT 2A

OXIDE	WT %
SI02	37.33
AL2O3	19.45
TiO2	0.00
FE2O3	1.98
FE0	27.16
MGO	1.63
MNO	0.45
CAO	10.76
NA2O	0.00
K2O	0.00
TOTAL	98.76

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.019
AL	
AL	1.855
TI	
FE3+	0.121
FE2+	1.837
MG	0.197
MN	0.031
CA	0.933
NA	
K	

ENDMEMBERS

% PYROPE	6.49
% ALM	60.54
% SPSS	1.01
% ANDR	5.96
% GROSS	24.78
% M/M+F	9.68

PC10 GT2B

OXIDE	WT %
SI02	37.34
AL2O3	19.17
TiO2	0.00
FE2O3	2.29
FE0	27.40
MGO	1.77
MNO	0.74
CAO	10.09
NA2O	0.00
K2O	0.00
TOTAL	98.79

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.024
AL	
AL	1.830
TI	
FE3+	0.140
FE2+	1.856
MG	0.214
MN	0.050
CA	0.876
NA	
K	

ENDMEMBERS

% PYROPE	7.03
% ALM	61.00
% SPSS	1.66
% ANDR	6.88
% GROSS	21.92
% M/M+F	10.34

PC10 GT 2C

OXIDE	WT %
SI02	37.28--
AL203	20.00
Ti02	0.00
FE203	2.09
FE0	27.94
MGO	1.73
MNO	0.61
CA0	10.49
NA20	0.00
K20	0.00
TOTAL	100.14

RECALCULATED ON
BASIS OF 12.0 (0)

SI	2.982
AL	0.018
AL	1.869
TI	
FE3+	0.126
FE2+	1.870
MG	0.206
MN	0.041
CA	0.899
NA	
K	

ENDMEMBERS

% PYROPE	6.79
% ALM	61.61
% SPSS	1.37
% ANDR	6.21
% GROSS	23.42
% M/M+F	9.93

PC10 GT RIM 2B

OXIDE	WT %
SI02	37.61
AL203	20.10
Ti02	0.00
FE203	1.31
FE0	28.28
MGO	2.21
MNO	0.26
CA0	9.51
NA20	0.00
K20	0.00
TOTAL	99.28 %

RECALCULATED ON
BASIS OF 12.0 (0)

SI	3.017
AL	
AL	1.901
TI	
FE3+	0.079
FE2+	1.897
MG	0.265
MN	0.018
CA	0.817
NA	
K	

ENDMEMBERS

% PYROPE	8.74
% ALM	62.68
% SPSS	0.59
% ANDR	3.92
% GROSS	23.08
% M/M+F	12.24

LM12 GT CORE 1

OXIDE	WT %
SI02	37.35
AL203	19.05
Ti02	0.00
FE203	1.86
FE0	24.10
MGO	1.68
MNO	2.47
CA0	11.02
NA20	0.00
K20	0.00
TOTAL	97.53 %

RECALCULATED ON
BASIS OF 12.0 (0)

SI	3.048
AL	
AL	1.833
TI	
FE3+	0.114
FE2+	1.645
MG	0.205
MN	0.170
CA	0.963
NA	
K	

ENDMEMBERS

% PYROPE	6.08
% ALM	53.69
% SPSS	5.56
% ANDR	5.59
% GROSS	25.85
% M/M+F	11.07

LM12 GT RIM 1		TUD5 GT 1 CORE		TUD5 GT 1 RIM A	
OXIDE	WT %	OXIDE	WT %	OXIDE	WT %
SI02	37.20	SI02	36.71	SI02	36.39
AL2O3	19.52	AL2O3	19.88	AL2O3	20.23
TI02	0.00	TI02	0.00	TI02	0.00
FE2O3	1.49	FE2O3	2.60	FE2O3	2.02
FEO	25.99	FEO	27.17	FEO	28.95
MGO	1.91	MGO	1.24	MGO	1.91
MNO	0.72	MNO	6.05	MNO	1.62
CAO	10.74	CAO	7.50	CAO	8.50
NA2O	0.00	NA2O	0.00	NA2O	0.00
K2O	0.00	K2O	0.00	K2O	0.00
TOTAL	97.58	TOTAL	101.15	TOTAL	99.62 ✓

RECALCULATED ON BASIS OF 12.0 (O) RECALCULATED ON BASIS OF 12.0 (O)

SI	3.030	SI	2.943
AL		AL	0.057
AL	1.875	AL	1.872
TI		TI	
FE3+	0.091	FE3+	0.123
FE2+	1.771	FE2+	1.958
MG	0.232	MG	0.230
MN	0.050	MN	0.111
CA	0.938	CA	0.737
NA		NA	
K		K	

ENDMEMBERS		ENDMEMBERS		ENDMEMBERS	
% PYROPE	7.64	% PYROPE	4.83	% PYROPE	7.43
% ALN	58.20	% ALM	59.24	% ALM	63.29
% SPESS	1.64	% SPESS	13.36	% SPESS	3.58
% ANDR	4.50	% ANDR	7.65	% ANDR	5.90
% GROSS	26.33	% GROSS	13.31	% GROSS	17.86
% M/M+F	11.61	% M/M+F	7.53	% M/M+F	11.51

TUD5 GT1 RIM B

OXIDE	WT %
SI02	36.04
AL2O3	19.71
TI02	0.00
FE2O3	2.27
FE0	27.73
MGO	2.01
MNO	1.75
CAO	8.74
NA2O	0.00
K2O	0.00
TOTAL	98.24

RECALCULATED ON
BASIS OF 12.0 (O)

SI	2.952
AL	0.048
AL	1.854
TI	
FE3+	0.140
FE2+	1.899
MG	0.245
MN	0.121
CA	0.767
NA	
K	

ENDMEMBERS

% PYROPE	7.95
% ALM	61.62
% SPSS	3.93
% ANDR	6.80
% GROSS	18.09
% M/M+F	11.15

TUD5 GT CORE 2

OXIDE	WT %
SI02	34.90
AL2O3	19.58
TI02	0.00
FE2O3	2.87
FE0	27.23
MGO	1.50
MNO	3.17
CAO	8.59
NA2O	0.00
K2O	0.00
TOTAL	97.84

RECALCULATED ON
BASIS OF 12.0 (O)

SI	2.897
AL	0.103
AL	1.813
TI	
FE3+	0.179
FE2+	1.890
MG	0.186
MN	0.223
CA	0.764
NA	
K	

ENDMEMBERS

% PYROPE	5.85
% ALM	59.60
% SPSS	7.03
% ANDR	8.47
% GROSS	15.62
% M/M+F	8.94

PM1 GT CORE 1

OXIDE	WT %
SI02	37.64
AL2O3	20.61
TI02	0.00
FE2O3	0.69
FE0	11.47
MGO	0.71
MNO	17.42
CAO	11.24
NA2O	0.00
K2O	0.00
TOTAL	99.77

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.012
AL	
AL	1.945
TI	
FE3+	0.041
FE2+	0.768
MG	0.084
MN	1.181
CA	0.964
NA	
K	

ENDMEMBERS

% PYROPE	2.79
% ALM	25.45
% SPSS	39.12
% ANDR	2.05
% GROSS	29.88
% M/M+F	9.89

PM1 GT1 RIM

OXIDE	WT %
SI02	38.15
AL2O3	21.39
TI02	0.00
FE2O3	0.00
FE0	28.66
MGO	4.08
MNO	1.26
CA0	6.28
NA2O	0.00
K2O	0.00
TOTAL	99.82

PM1 GT2 RIM

OXIDE	WT %
SI02	37.90
AL2O3	21.47
TI02	0.00
FE2O3	0.62
FE0	27.53
MGO	3.21
MNO	1.45
CA0	8.76
NA2O	0.00
K2O	0.00
TOTAL	100.94

PM1 GT2 CORE

OXIDE	WT %
SI02	37.94
AL2O3	21.17
TI02	0.00
FE2O3	0.40
FE0	15.51
MGO	1.30
MNO	11.51
CA0	12.33
NA2O	0.00
K2O	0.00
TOTAL	100.22

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.012
AL	
AL	1.991
TI	
FE3+	
FE2+	1.893
MG	0.480
MN	0.084
CA	0.531
NA	
K	

ENDMEMBERS

% PYROPE	16.00
% ALM	63.06
% SPSS	2.81
% ANDR	0.00
% GROSS	17.69
% M/M+F	20.24

RECALCULATED ON
BASIS OF 12.0 (O)

SI	2.975
AL	0.025
AL	1.962
TI	
FE3+	0.037
FE2+	1.807
MG	0.375
MN	0.096
CA	0.730
NA	
K	

ENDMEMBERS

% PYROPE	12.34
% ALM	59.43
% SPSS	3.17
% ANDR	1.81
% GROSS	22.40
% M/M+F	17.20

RECALCULATED ON
BASIS OF 12.0 (O)

SI	3.001
AL	
AL	1.974
TI	
FE3+	0.024
FE2+	1.020
MG	0.160
MN	0.771
CA	1.045
NA	
K	

ENDMEMBERS

% PYROPE	5.32
% ALM	34.15
% SPSS	25.60
% ANDR	1.20
% GROSS	33.58
% M/M+F	13.48

KB1 GLAUC 1

OXIDE	WT %
SI02	56.23
AL203	9.10
TI02	0.00
FE203	5.44
FE0	11.41
MGO	7.72
MNO	0.05
CAO	0.34
NA2O	6.69
K2O	0.03
TOTAL	97.01

RECALCULATED ON
BASIS OF 23.0 (0)

SI	7.952
AL	0.048
AL	1.468
TI	
FE3+	0.579
FE2+	1.350
MG	1.628
MN	0.006
CA	0.051
NA	1.835
K	0.005

ENDMEMBERS

% TREM 2.55

KB1 GLAUC CORE 2

OXIDE	WT %
SI02	54.65
AL203	8.84
TI02	0.00
FE203	5.87
FE0	12.33
MGO	6.94
MNO	0.03
CAO	0.35
NA2O	7.19
K2O	0.04
TOTAL	96.23

RECALCULATED ON
BASIS OF 23.0 (0)

SI	7.875
AL	0.125
AL	1.377
TI	
FE3+	0.637
FE2+	1.486
MG	1.490
MN	0.004
CA	0.054
NA	2.009
K	0.007

ENDMEMBERS

% TREM 2.71

KB1 GLAUC 3

OXIDE	WT %
SI02	56.92
AL203	7.75
TI02	0.00
FE203	5.81
FE0	12.19
MGO	7.27
MNO	0.08
CAO	0.35
NA2O	6.83
K2O	0.03
TOTAL	97.22

RECALCULATED ON
BASIS OF 23.0 (0)

SI	8.070
AL	
AL	1.296
TI	
FE3+	0.620
FE2+	1.446
MG	1.536
MN	0.010
CA	0.053
NA	1.877
K	0.005

ENDMEMBERS

% TREM 2.66

PC10 GLAUC 2

OXIDE	WT %
SI02	54.40
AL203	9.06
TI02	0.00
FE203	4.09
FE0	8.58
MGO	9.52
MNO	0.04
CA0	0.94
NA20	6.76
K20	0.04
TOTAL	93.41

RECALCULATED ON
BASIS OF 23.0 (0)

SI	7.898
AL	0.102
AL	1.450
TI	
FE3+	0.446
FE2+	1.042
MG	2.060
MN	0.004
CA	0.146
NA	1.902
K	0.007

ENDMEMBERS

% TREM 7.28

PC10 GLAUC 3

OXIDE	WT %
SI02	56.69
AL203	10.84
TI02	0.00
FE203	3.92
FE0	8.24
MGO	9.63
MNO	0.13
CA0	1.15
NA20	6.72
K20	0.05
TOTAL	97.38

RECALCULATED ON
BASIS OF 23.0 (0)

SI	7.849
AL	0.151
AL	1.619
TI	
FE3+	0.409
FE2+	0.904
MG	1.907
MN	0.016
CA	0.171
NA	1.803
K	0.009

ENDMEMBERS

% TREM 8.55

PC10 GLAUC 4

OXIDE	WT %
SI02	55.23
AL203	10.33
TI02	0.00
FE203	4.97
FE0	10.43
MGO	8.75
MNO	0.06
CA0	1.92
NA20	6.53
K20	0.05
TOTAL	98.27

RECALCULATED ON
BASIS OF 23.0 (0)

SI	7.718
AL	0.282
AL	1.421
TI	
FE3+	0.522
FE2+	1.219
MG	1.822
MN	0.008
CA	0.288
NA	1.770
K	0.009

ENDMEMBERS

% TREM 14.38

PC10 GLAUC 5		
OXIDE	WT %	
SI02	57.65	
AL203	11.14	
TI02	0.00	
FE203	4.02	
FE0	8.44	
MGO	9.49	
MNO	0.05	
CA0	0.67	
NA20	7.03	
K20	0.04	
TOTAL	98.54	

RECALCULATED ON BASIS OF 23.0 (0)		
SI	7.877	
AL	0.123	
AL	1.670	
TI		
FE3+	0.413	
FE2+	0.964	
MG	1.933	
MN	0.006	
CA	0.099	
NA	1.862	
K	0.006	

ENDMEMBERS

% TREM 4.93

PC10 GLAUC 6		
OXIDE	WT %	
SI02	49.56	
AL203	8.77	
TI02	0.00	
FE203	5.96	
FE0	12.51	
MGO	9.38	
MNO	0.09	
CA0	7.63	
NA20	3.76	
K20	0.23	
TOTAL	97.90	

RECALCULATED ON BASIS OF 23.0 (0)		
SI	7.222	
AL	0.778	
AL	0.729	
TI		
FE3+	0.653	
FE2+	1.524	
MG	2.037	
MN	0.011	
CA	1.191	
NA	1.063	
K	0.043	

ENDMEMBERS

% TREM 40.73

PC10 GLAUC 7		
OXIDE	WT %	
SI02	50.30	
AL203	7.76	
TI02	0.00	
FE203	5.78	
FE0	12.14	
MGO	9.88	
MNO	0.11	
CA0	8.39	
NA20	3.50	
K20	0.31	
TOTAL	98.17	

RECALCULATED ON BASIS OF 23.0 (0)		
SI	7.303	
AL	0.697	
AL	0.632	
TI		
FE3+	0.632	
FE2+	1.474	
MG	2.139	
MN	0.013	
CA	1.300	
NA	0.986	
K	0.057	

ENDMEMBERS

% TREM 42.77

TUD5 GLAUC 2		
OXIDE	WT %	
SI02	53.88	
AL203	8.69	
TI02	0.00	
FE203	5.45	
FFO	11.43	
MGO	7.64	
MNO	0.13	
CAO	0.74	
NA2O	6.87	
K2O	0.04	
TOTAL	94.85	

RECALCULATED ON BASIS OF 23.0 (O)		
SI	7.854	
AL	0.146	
AL	1.347	
TI		
FE3+	0.597	
FE2+	1.394	
MG	1.659	
MN	0.012	
CA	0.116	
NA	1.943	
K	0.008	

ENDMEMBERS

% TREM 5.81

TUD5 GLAUC 3 CORE		
OXIDE	WT %	
SI02	54.69	
AL203	8.88	
TI02	0.00	
FE203	4.88	
FLO	10.25	
MGO	8.59	
MNO	0.09	
CAO	0.58	
NA2O	7.14	
K2O	0.03	
TOTAL	95.14	

RECALCULATED ON BASIS OF 23.0 (O)		
SI	7.881	
AL	0.119	
AL	1.389	
TI		
FE3+	0.529	
FE2+	1.235	
MG	1.840	
MN	0.011	
CA	0.089	
NA	1.990	
K	0.000	

ENDMEMBERS

% TREM 4.47

TUD5 GLAUC 3 RIM		
OXIDE	WT %	
SI02	44.91	
AL203	8.59	
TI02	0.00	
FE203	6.28	
FE0	13.20	
MGO	10.29	
MNO	0.34	
CAO	8.15	
NA2O	3.40	
K2O	0.33	
TOTAL	95.48	

RECALCULATED ON BASIS OF 23.0 (O)		
SI	6.841	
AL	1.159	
AL	0.383	
TI		
FE3+	0.720	
FE2+	1.681	
MG	2.330	
MN	0.044	
CA	1.330	
NA	1.003	
K	0.003	

ENDMEMBERS

% TREM 46.71

TUDB BARROISITE 1		
OXIDE	WT %	
SI02	47.81	
AL203	6.72	
TI02	0.00	
FE203	6.38	
FE0	13.39	
MGO	12.28	
MNO	0.34	
CA0	8.85	
NA20	2.64	
K20	0.14	
TOTAL	98.53	

RECALCULATED ON BASIS OF 23.0 (0)		
SI	7.021	
AL	0.979	
AL	0.184	
TI		
FE3+	0.705	
FE2+	1.645	
MG	2.688	
MN	0.043	
CA	1.392	
NA	0.751	
K	0.027	
ENDMEMBERS		
X TREM	53.75	

PC5 GLAUC 1		
OXIDE	WT %	
SI02	55.06	
AL203	9.50	
TI02	0.00	
FE203	4.14	
FE0	8.68	
MGO	9.53	
MNO	0.07	
CA0	0.36	
NA20	7.61	
K20	0.03	
TOTAL	94.98	

RECALCULATED ON BASIS OF 23.0 (0)		
SI	7.871	
AL	0.129	
AL	1.472	
TI		
FE3+	0.445	
FE2+	1.030	
MG	2.031	
MN	0.008	
CA	0.055	
NA	2.110	
K	0.006	
ENDMEMBERS		
% TREM	2.73	

PC5 GA 1		
OXIDE	WT %	
SI02	46.89	
AL203	9.00	
TI02	0.00	
FE203	5.94	
FE0	12.46	
MGO	10.22	
MNO	0.15	
CA0	6.93	
NA20	4.60	
K20	0.21	
TOTAL	96.39	

RECALCULATED ON BASIS OF 23.0 (0)		
SI	6.994	
AL	1.006	
AL	0.577	
TI		
FE3+	0.666	
FE2+	1.555	
MG	2.272	
MN	0.018	
CA	1.100	
NA	1.332	
K	0.040	
ENDMEMBERS		
% TRFM	45.44	

PC5 GLAUC 2

OXIDE	WT %
SI02	56.29
AL2O3	9.31
TI02	0.00
FE2O3	4.30
FE0	9.04
MGO	9.57
MNO	0.06
CAO	0.40
NA2O	7.44
K2O	0.02
TOTAL	96.43

RECALCULATED ON
BASIS OF 23.0 (O)

SI	7.922
AL	0.078
AL	1.466
TI	
FE3+	0.456
FE2+	1.064
MG	2.008
MN	0.007
CA	0.061
NA	2.029
K	0.004

ENDMEMBERS

% TREM	3.03
--------	------

PC5 GA 2

OXIDE	WT %
SI02	42.86
AL2O3	8.45
TI02	0.00
FE2O3	5.93
FE0	12.44
MGO	10.02
MNO	0.16
CAO	7.53
NA2O	4.37
K2O	0.17
TOTAL	91.93

RECALCULATED ON
BASIS OF 23.0 (O)

SI	6.787
AL	1.213
AL	0.365
TI	
FE3+	0.706
FE2+	1.048
MG	2.365
MN	0.021
CA	1.277
NA	1.341
K	0.035

ENDMEMBERS

% TREM	47.31
--------	-------

PM1 GLAUC 1

OXIDE	WT %
SI02	56.43
AL2O3	11.08
TI02	0.00
FE2O3	3.09
FE0	6.49
MGO	11.27
MNO	0.07
CAO	0.94
NA2O	7.22
K2O	0.01
TOTAL	96.61

RECALCULATED ON
BASIS OF 23.0 (O)

SI	7.807
AL	0.193
AL	1.615
TI	
FE3+	0.322
FE2+	0.751
MG	2.324
MN	0.008
CA	0.139
NA	1.938
K	0.001

ENDMEMBERS

% TREM	6.94
--------	------

PN1 GLAUC 1 RTM

OXIDE	WT %
SI02	55.43
AL203	8.33
TI02	0.00
FE203	5.78
FE0	12.14
MGO	8.17
MNO	0.06
CAO	0.47
NA2O	6.90
K2O	0.04
TOTAL	97.34 ✓

RECALCULATED ON BASIS OF 23.0 (O)	
SI	7.883
AL	0.117
AL	1.280
TI	0.619
FE3+	1.444
FE2+	1.732
MG	0.008
MN	0.071
CA	1.904
NA	0.008
K	

ENDMEMBERS

% TREM 3.55

PN1 GLAUC 2 CORE

OXIDE	WT %
SI02	57.55
AL203	9.11
TI02	0.00
FE203	5.28
FE0	11.08
MGO	8.21
MNO	0.04
CAO	0.44
NA2O	7.15
K2O	0.02
TOTAL	98.88 ✓

RECALCULATED ON BASIS OF 23.0 (O)	
SI	7.968
AL	0.032
AL	1.456
TI	0.550
FE3+	1.283
FE2+	1.693
MG	0.005
MN	0.065
CA	1.919
NA	0.003
K	

ENDMEMBERS

% TREM 3.23

LM12 GLAUC 1

OXIDE	WT %
SI02	56.31
AL203	10.03
TI02	0.00
FE203	3.44
FE0	7.22
MGO	9.47
MNO	0.04
CAO	1.26
NA2O	7.15
K2O	0.03
TOTAL	94.94

RECALCULATED ON BASIS OF 23.0 (O)	
SI	7.959
AL	0.041
AL	1.630
TI	0.366
FE3+	0.854
FE2+	1.994
MG	0.005
MN	0.190
CA	1.960
NA	0.006
K	

ENDMEMBERS

% TREM 9.52

LM12 GLAUC 2			LM12 GLAUC 3			BIL3 ACTINOLITE 1		
OXIDE	WT %		OXIDE	WT %		OXIDE	WT %	
SI02	56.10		SI02	56.99		SI02	52.47	
AL203	10.08		AL203	9.85		AL203	4.42	
TI02	0.00		TI02	0.00		TI02	0.00	
FE203	3.36		FE203	3.36		FE203	3.23	
FE0	7.05		FE0	7.05		FE0	6.79	
MGO	10.54		MGO	10.05		MGO	16.07	
MNO	0.04		MNO	0.01		MNO	0.29	
CAU	0.76		CAU	1.87		CAU	10.94	
NA20	7.24		NA20	6.90		NA20	1.94	
K20	0.04		K20	0.02		K20	0.19	
TOTAL	95.21		TOTAL	96.10		TOTAL	96.35	

RECALCULATED ON BASIS OF 23.0 (O)			RECALCULATED ON BASIS OF 23.0 (O)			RECALCULATED ON BASIS OF 23.0 (O)		
SI	7.901		SI	7.955		SI	7.542	
AL	0.099		AL	0.045		AL	0.458	
AL	1.575		AL	1.576		AL	0.291	
TI			TI			TI		
FE3+	0.356		FE3+	0.353		FE3+	0.350	
FE2+	0.830		FE2+	0.823		FE2+	0.816	
MG	2.211		MG	2.090		MG	3.443	
MN	0.005		MN	0.002		MN	0.036	
CA	0.115		CA	0.279		CA	1.684	
NA	1.977		NA	1.867		NA	0.541	
K	0.007		K	0.004		K	0.035	

ENDMEMBERS			ENDMEMBERS			ENDMEMBERS		
% TREM	5.76		% TREM	13.96		% TREM	08.87	

BIL3 ACTINOLITE 2		LM12 CHL 1	
OXIDE	WT %	OXIDE	WT %
SI02	50.87	SI02	42.67
AL2O3	4.65	AL2O3	4.45
TI02	0.00	TI02	0.00
FE2O3	3.36	FE2O3	4.70
FE0	7.05	FE0	9.88
MGO	16.43	MGO	13.09
MNO	0.31	MNO	0.29
CA0	11.20	CA0	8.54
NA20	1.67	NA20	0.00
K20	0.20	K20	0.00
TOTAL	95.74	TOTAL	83.63

LM1 ACTINOLITE			
OXIDE	WT %	OXIDE	WT %
SI02	49.96		
AL2O3	5.26		
TI02	0.00		
FE2O3	2.99		
FE0	6.28		
MGO	13.97		
MNO	0.32		
CA0	9.91		
NA20	2.09		
K20	0.16		
TOTAL	90.95		

LM12 CHL 1

RECALCULATED ON BASIS OF 23.0 (0)		RECALCULATED ON BASIS OF 23.0 (0)	
SI	7.398	SI	7.233
AL	0.602	AL	0.767
AL	0.195	AL	0.122
TI		TI	
FE3+	0.367	FE3+	0.600
FE2+	0.858	FE2+	1.400
MG	3.502	MG	3.307
MN	0.038	MN	0.042
CA	1.746	CA	1.500
NA	0.472	NA	
K	0.037	K	
ENDMEMBERS		ENDMEMBERS	
% TREM	71.24	% TREM	66.14

RECALCULATED ON BASIS OF 23.0 (0)		RECALCULATED ON BASIS OF 23.0 (0)	
SI	7.576	SI	7.233
AL	0.424	AL	0.767
AL	0.515	AL	0.122
TI		TI	
FE3+	0.341	FE3+	0.600
FE2+	0.797	FE2+	1.400
MG	3.158	MG	3.307
MN	0.041	MN	0.042
CA	1.609	CA	1.500
NA	0.614	NA	
K	0.031	K	
ENDMEMBERS		ENDMEMBERS	
% TREM	63.15	% TREM	66.14

RECALCULATED ON BASIS OF 23.0 (0)		RECALCULATED ON BASIS OF 23.0 (0)	
SI	7.576	SI	7.233
AL	0.424	AL	0.767
AL	0.515	AL	0.122
TI		TI	
FE3+	0.341	FE3+	0.600
FE2+	0.797	FE2+	1.400
MG	3.158	MG	3.307
MN	0.041	MN	0.042
CA	1.609	CA	1.500
NA	0.614	NA	
K	0.031	K	
ENDMEMBERS		ENDMEMBERS	
% TREM	63.15	% TREM	66.14

TUD5 PHENG 1

OXIDE	WT %
SI02	47.96
AL203	26.54
TI02	0.00
FE203	0.00
FE0	4.56
MGO	2.95
MNO	0.04
CA0	0.03
NA20	0.55
K20	10.07
TOTAL	92.64

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.679
AL	1.321
AL	3.028
TI	
FE3+	
FE2+	0.531
MG	0.611
MN	0.004
CA	0.004
NA	0.148
K	1.789

ENDMEMBERS

% MUSC	61.02
% PARAG	5.04
% CELAD	33.94
% M/M+F	53.53

PM1 PARAGONITE

OXIDE	WT %
SI02	46.18
AL203	39.12
TI02	0.00
FE203	0.00
FE0	0.74
MGO	0.49
MNO	0.02
CA0	0.12
NA20	6.16
K20	0.70
TOTAL	93.55

RECALCULATED ON
BASIS OF 22.0 (0)

SI	5.995
AL	2.005
AL	3.982
TI	
FE3+	
FE2+	0.080
MG	0.096
MN	0.003
CA	0.017
NA	1.550
K	0.110

ENDMEMBERS

% MUSC	5.81
% PARAG	77.49
% CELAD	0.00
% M/M+F	54.35

PC10 WM 1

OXIDE	WT %
SI02	48.53
AL203	24.35
TI02	0.00
FE203	0.00
FE0	3.39
MGO	2.89
MNO	0.03
CA0	0.02
NA20	0.13
K20	9.49
TOTAL	88.84

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.956
AL	1.044
AL	3.071
TI	
FE3+	
FE2+	0.406
MG	0.617
MN	0.004
CA	0.004
NA	0.036
K	1.735

ENDMEMBERS

% MUSC	51.17
% PARAG	1.05
% CELAD	46.46
% M/M+F	60.29

PN1 PHENGITE

OXIDE	WT %
SI02	48.15
AL203	25.21
TI02	0.00
FE203	0.00
FE0	5.33
MGO	3.20
MNO	0.03
CA0	0.03
NA20	0.35
K20	10.46
TOTAL	92.78

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.739
AL	1.261
AL	2.898
TI	
FE3+	0.624
FE2+	0.668
MG	0.004
MN	0.005
CA	0.090
NA	1.868
K	

ENDMEMBERS

% MUSC	50.97
% PARAG	3.08
% CELAD	36.95
% M/M+F	51.69

PC10 WM 2

OXIDE	WT %
SI02	49.40
AL203	25.21
TI02	0.00
FE203	0.00
FE0	3.65
MGO	3.51
MNO	0.03
CA0	0.04
NA20	0.31
K20	7.79
TOTAL	89.95

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.922
AL	1.078
AL	3.087
TI	
FE3+	
FE2+	0.428
MG	0.733
MN	0.003
CA	0.006
NA	0.083
K	1.393

ENDMEMBERS

% MUSC	50.87
% PARAG	3.03
% CELAD	45.67
% M/M+F	63.15

PC10 PHENG 1 EDS

OXIDE	WT %
SI02	48.92
AL203	24.34
TI02	0.39
FE203	0.00
FE0	3.41
MGO	3.23
MNO	0.00
CA0	0.18
NA20	0.00
K20	10.24
TOTAL	90.70

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.905
AL	1.095
AL	2.956
TI	0.041
FE3+	
FE2+	0.402
MG	0.680
MN	
CA	0.027
NA	
K	1.044

ENDMEMBERS

% MUSC	54.73
% PARAG	0.00
% CELAD	45.27
% M/M+F	62.83

PC10 PHENG 2 EDS

OXIDE	WT %
SI02	48.72
AL2O3	25.00
TI02	0.27
FE2O3	0.00
FE0	4.32
MGO	2.84
MNO	0.00
CAO	0.28
NA2O	0.00
K2O	10.27
TOTAL	91.70

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.838
AL	1.162
AL	2.973
TI	0.029
FE3+	0.507
FE2+	0.594
MG	0.043
MN	0.043
CA	0.043
NA	0.043
K	1.838

ENDMEMBERS

% MUSC	58.12
% PARAG	0.00
% CELAD	41.88
% M/M+F	53.95

KB1 PARA 1 EDS

OXIDE	WT %
SI02	45.75
AL2O3	33.54
TI02	0.23
FE2O3	0.00
FE0	3.33
MGO	1.29
MNO	0.00
CAO	0.13
NA2O	4.48
K2O	1.25
TOTAL	90.01

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.246
AL	1.754
AL	3.043
TI	0.024
FE3+	0.380
FE2+	0.263
MG	0.019
MN	0.019
CA	1.187
NA	0.218
K	0.218

LNDMEMBERS

% MUSC	10.92
% PARAG	59.36
% CELAD	12.30
% M/M+F	40.87

KB1 MIC 1 EDS

OXIDE	WT %
SI02	45.58
AL2O3	28.78
TI02	0.15
FE2O3	0.00
FE0	4.10
MGO	1.43
MNO	0.00
CAO	0.00
NA2O	0.00
K2O	8.65
TOTAL	88.67

RECALCULATED ON
BASIS OF 22.0 (0)

SI	6.539
AL	1.401
AL	3.407
TI	0.016
FE3+	0.492
FE2+	0.305
MG	0.000
MN	0.000
CA	0.000
NA	0.000
K	1.583

ENDMEMBERS

% MUSC	73.04
% PARAG	0.00
% CELAD	26.96
% M/M+F	38.31

PM1 CLINOZOISITE

PM1 FP 1

BIL3 EP 1

OXIDE	WT %
SI02	38.04
AL2O3	27.18
TI02	0.00
FE2O3	7.58
FE0	0.00
MGO	0.23
MNO	0.23
CAO	23.51
NA2O	0.00
K2O	0.00
TOTAL	96.78

OXIDE	WT %
SI02	37.23
AL2O3	22.64
TI02	0.00
FE2O3	12.73
FE0	0.00
MGO	0.17
MNO	0.12
CAO	23.13
NA2O	0.00
K2O	0.00
TOTAL	96.02

OXIDE	WT %
SI02	36.77
AL2O3	22.88
TI02	0.00
FE2O3	10.44
FE0	0.00
MGO	0.17
MNO	0.19
CAO	22.98
NA2O	0.00
K2O	0.00
TOTAL	93.42

RECALCULATED ON
BASIS OF 12.5 (O)

SI	3.001
AL	
AL	2.528
TI	
FE3+	0.450
FE2+	
MG	0.027
MN	0.015
CA	1.988
NA	
K	

ENDMEMBERS

% EP	15.00
% CZ	84.27

RECALCULATED ON
BASIS OF 12.5 (O)

SI	3.022
AL	
AL	2.166
TI	
FE3+	0.778
FE2+	
MG	0.021
MN	0.008
CA	2.011
NA	
K	

ENDMEMBERS

% EP	25.92
% CZ	72.21

RECALCULATED ON
BASIS OF 12.5 (O)

SI	3.047
AL	
AL	2.236
TI	
FE3+	0.651
FE2+	
MG	0.022
MN	0.013
CA	2.041
NA	
K	

ENDMEMBERS

% EP	21.70
% CZ	74.52

APPENDIX IVGRAPHICAL PRESENTATION OF MINERAL ANALYSES

Zoning paths in Groix garnets	IV.1
Triangular composition diagram for clinopyroxenes	IV.2
Partition of Fe/Mg between co-existing garnet and clinopyroxene	IV.3
Composition of glaucophanes	IV.4
Projection of co-existing mineral compositions from epidote in the subsystem $\text{CaO-Al}_2\text{O}_3\text{-FeO-MgO}$	IV.5
Compositions of micas	IV.6

Fig. IV.1

Zoning paths in
Groix garnets

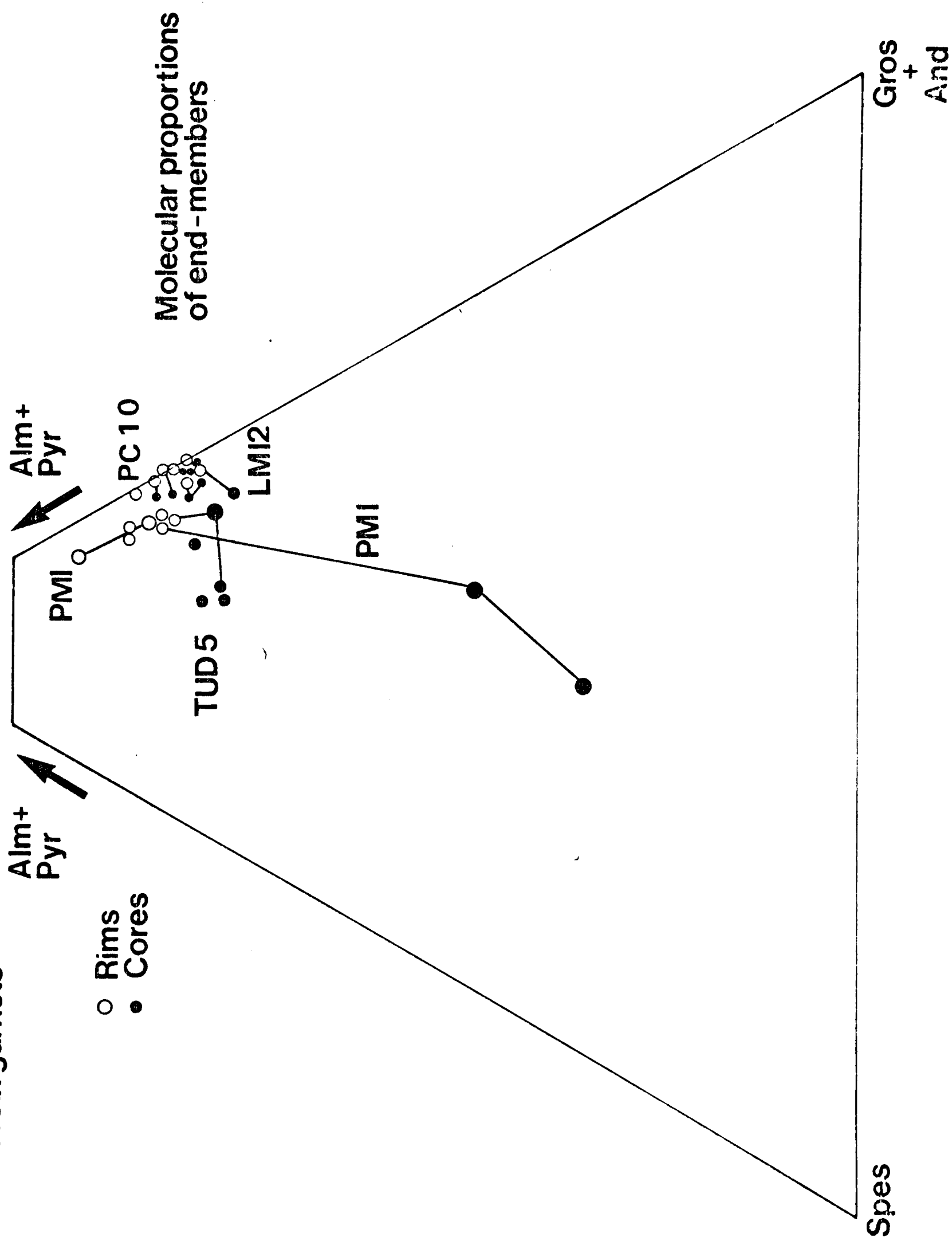


Fig. IV.2

Triangular composition diagram showing proportions of end-members in pyroxenes from the Ile de Groix.

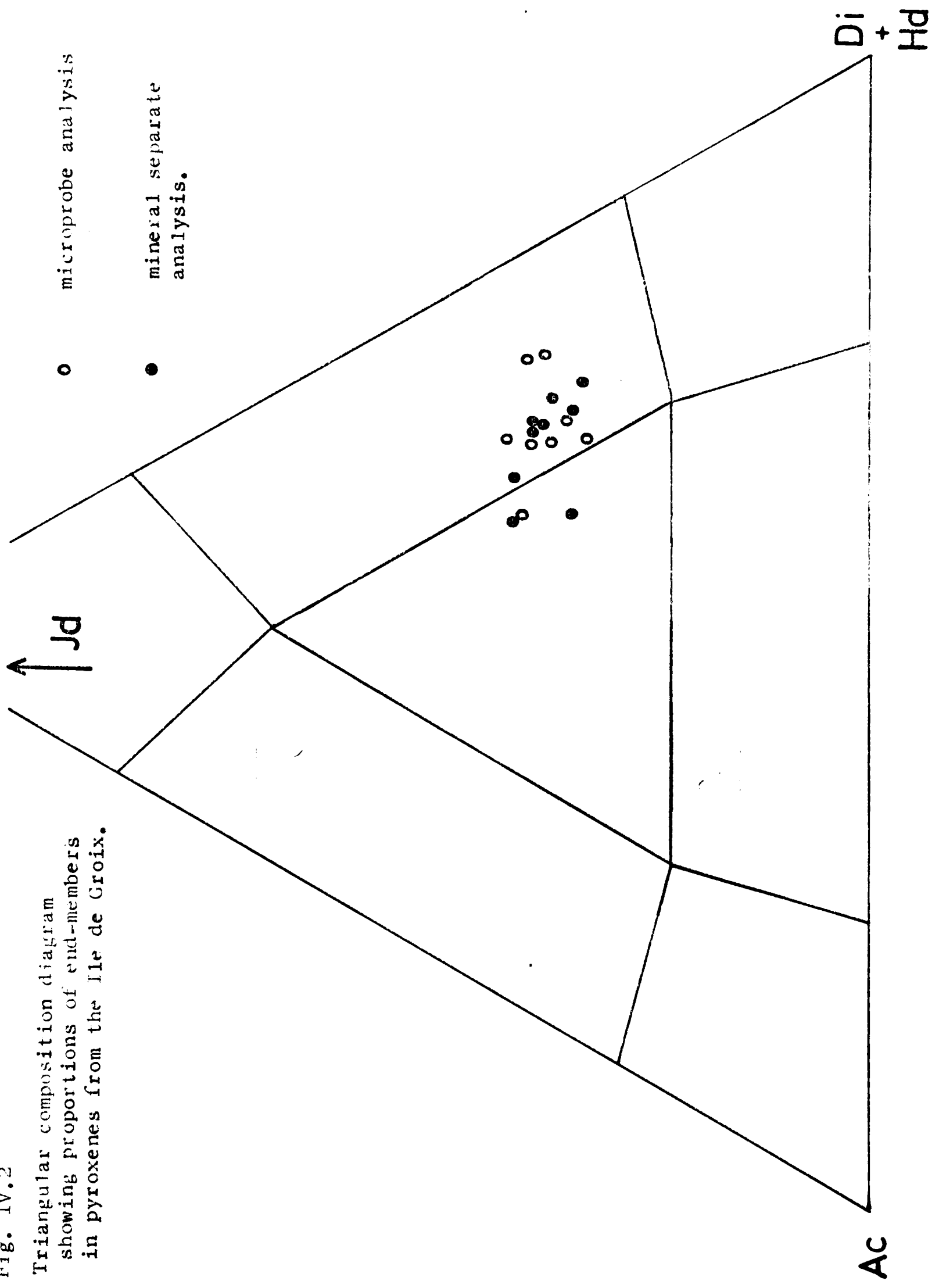
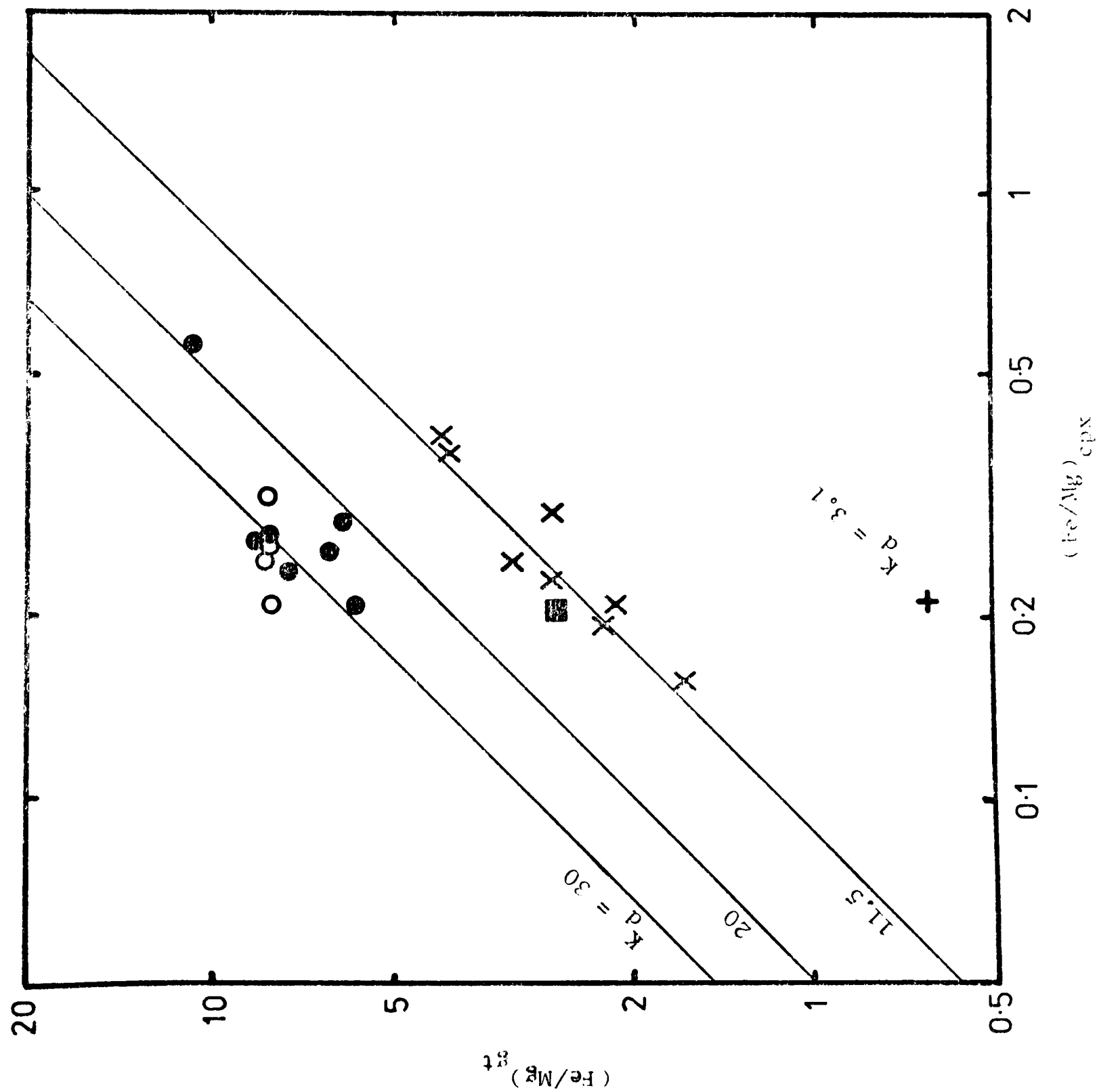


Fig. IV.3

Partition of Fe/Mg between coexisting
garnet and clinopyroxene.



Groix glaucophanic eclogites:

- mineral separates
- microprobe spots

■ Californian exotic eclogitic blocks, Valley Ford, Coleman et al., (1965).

× Eclogites from amphibolitic gneisses near Nantes, France Velde, (1970).

+ Knockormal eclogite, Girvan, Bloxham and Allen, (1958).

Fig. IV.4
Composition of Groix glaucophanes, with Fe^{3+} recalculated on the basis $(\text{Fe}^{3+} + \text{Al}^{\text{vi}}) = \text{Na}^{+}$ in the riebeckite molecule.

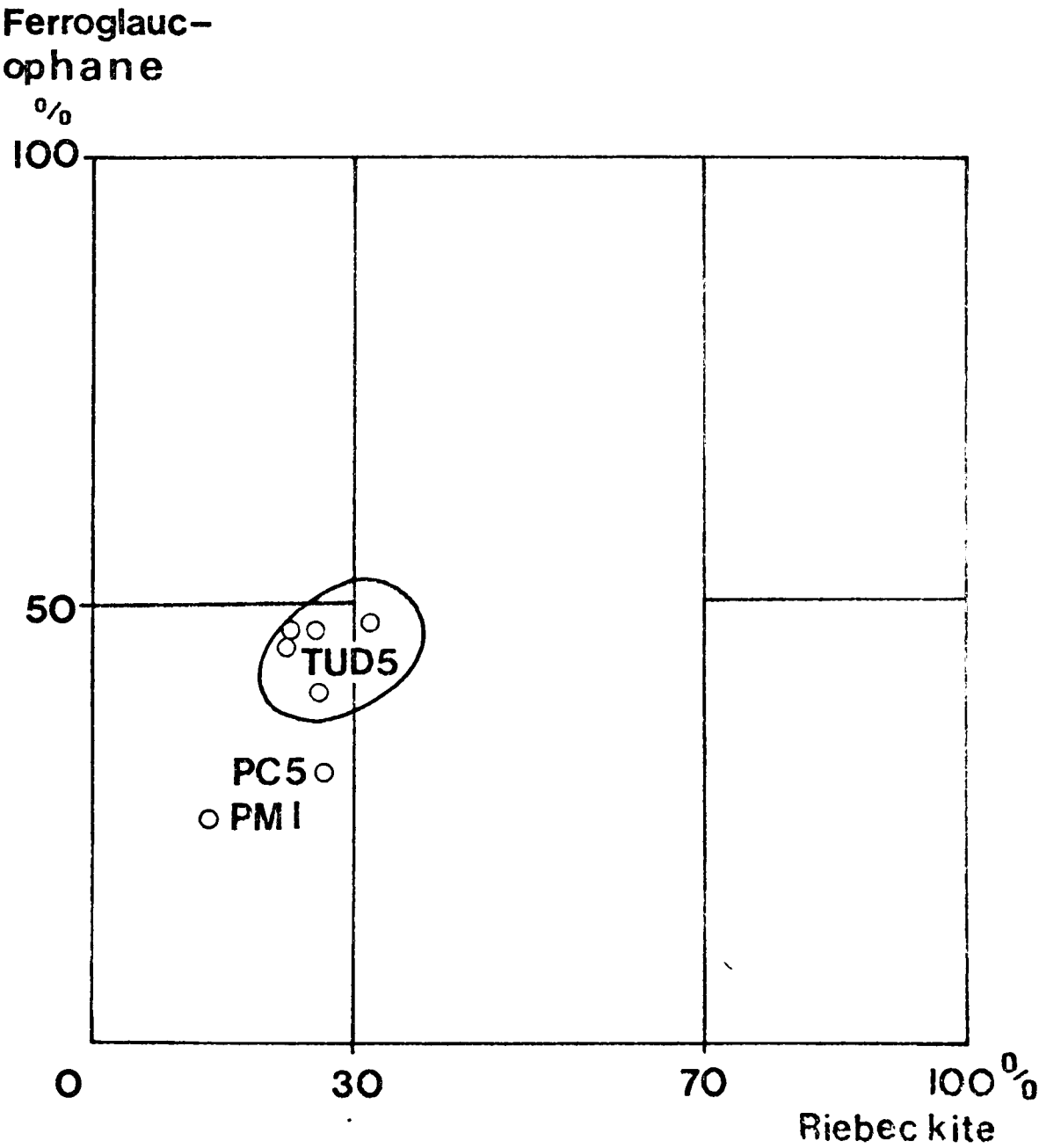
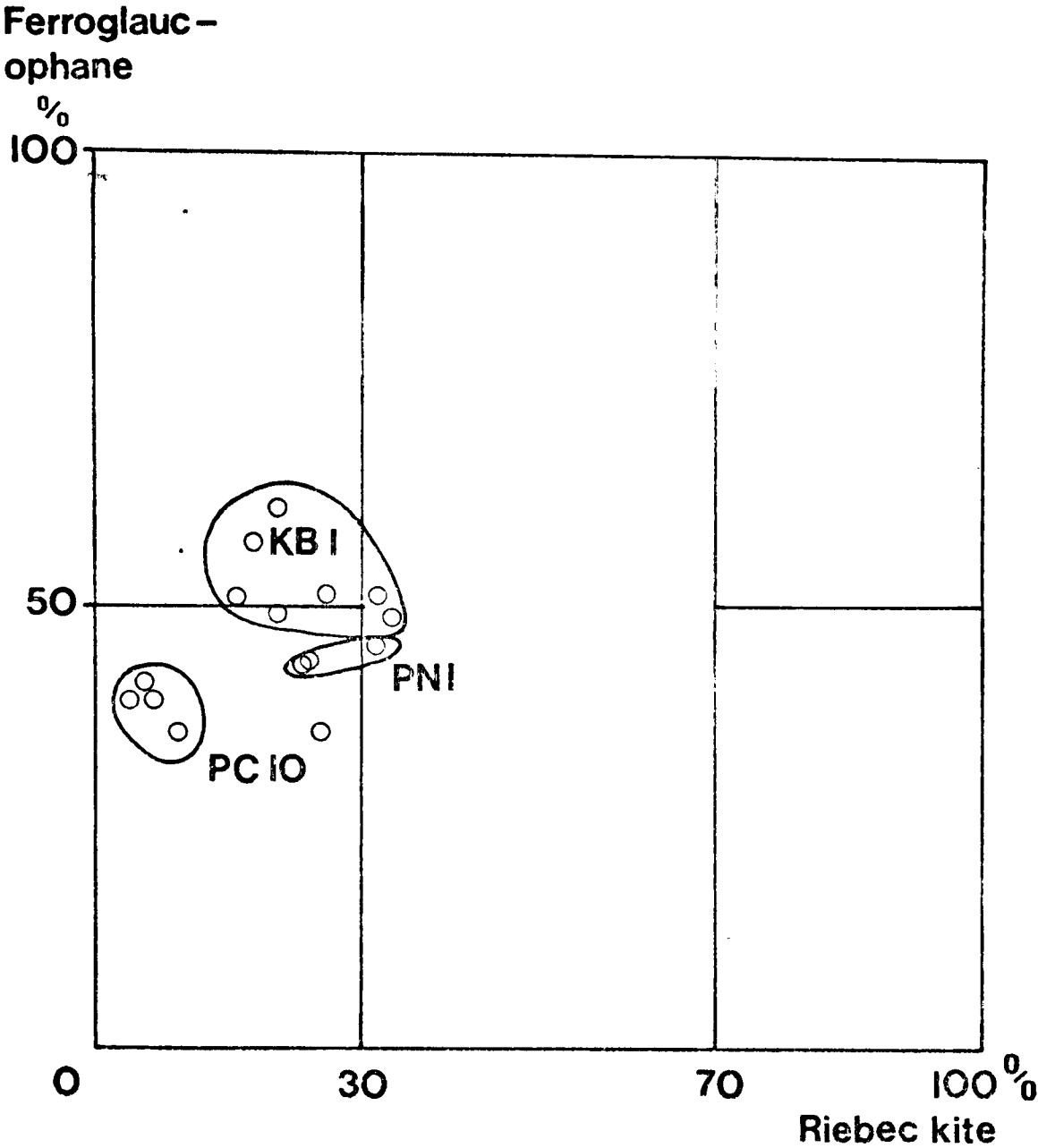


Fig. IV.5

Glauconophane schist mineral compositions projected from epidote in the subsystem $\text{CaO-Al}_2\text{O}_3\text{-FeO-MgO}$ onto the plane containing FeO-MgO and parallel to $\text{CaO-Al}_2\text{O}_3$. The vertical axis is calculated as $(\text{Al}_2\text{O}_3 - \text{Na}_2\text{O} - \frac{3}{4}\text{CaO} + \text{FeO} + \text{MgO}) / (\text{FeO} + \text{MgO})$. Molecular proportions of oxides.

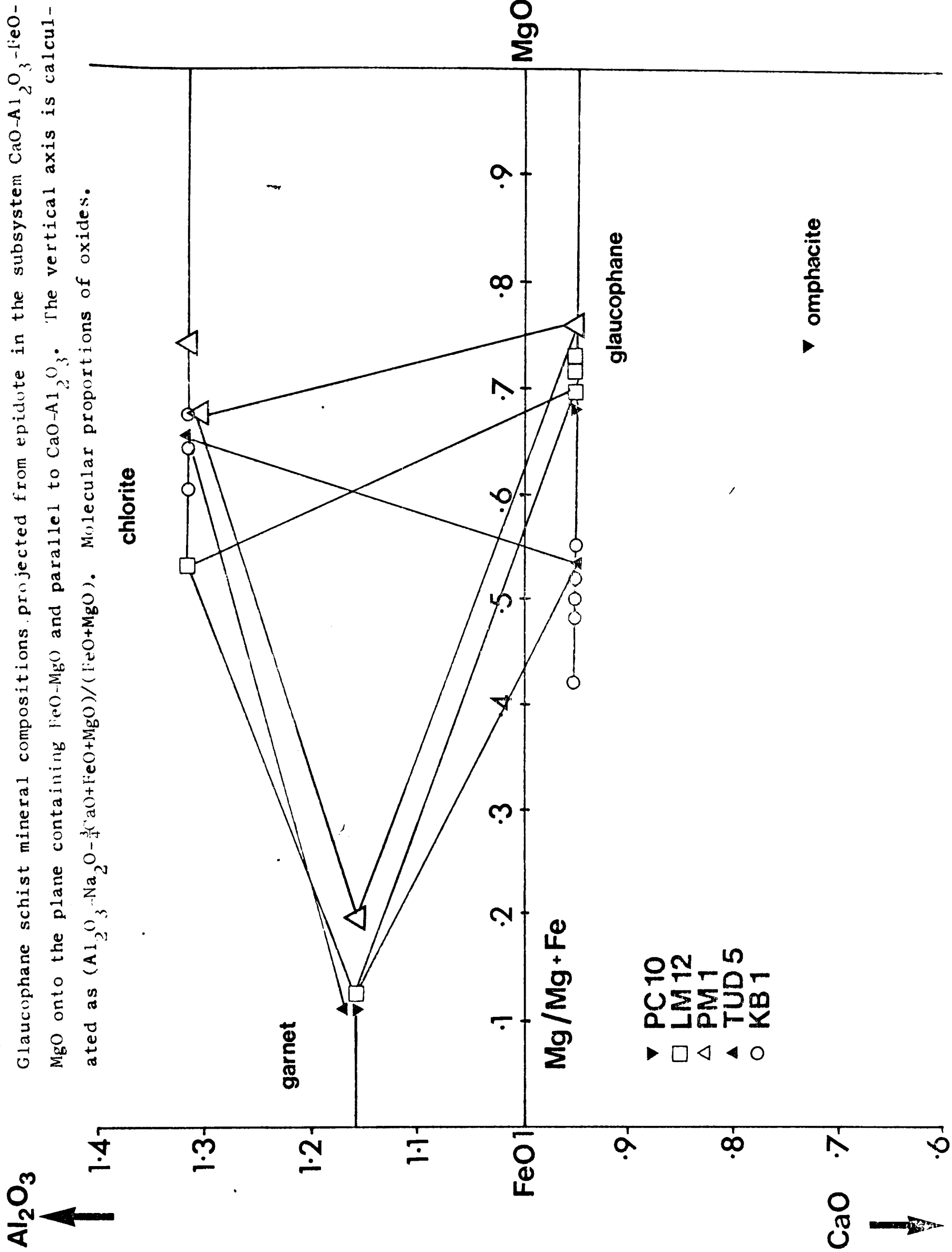


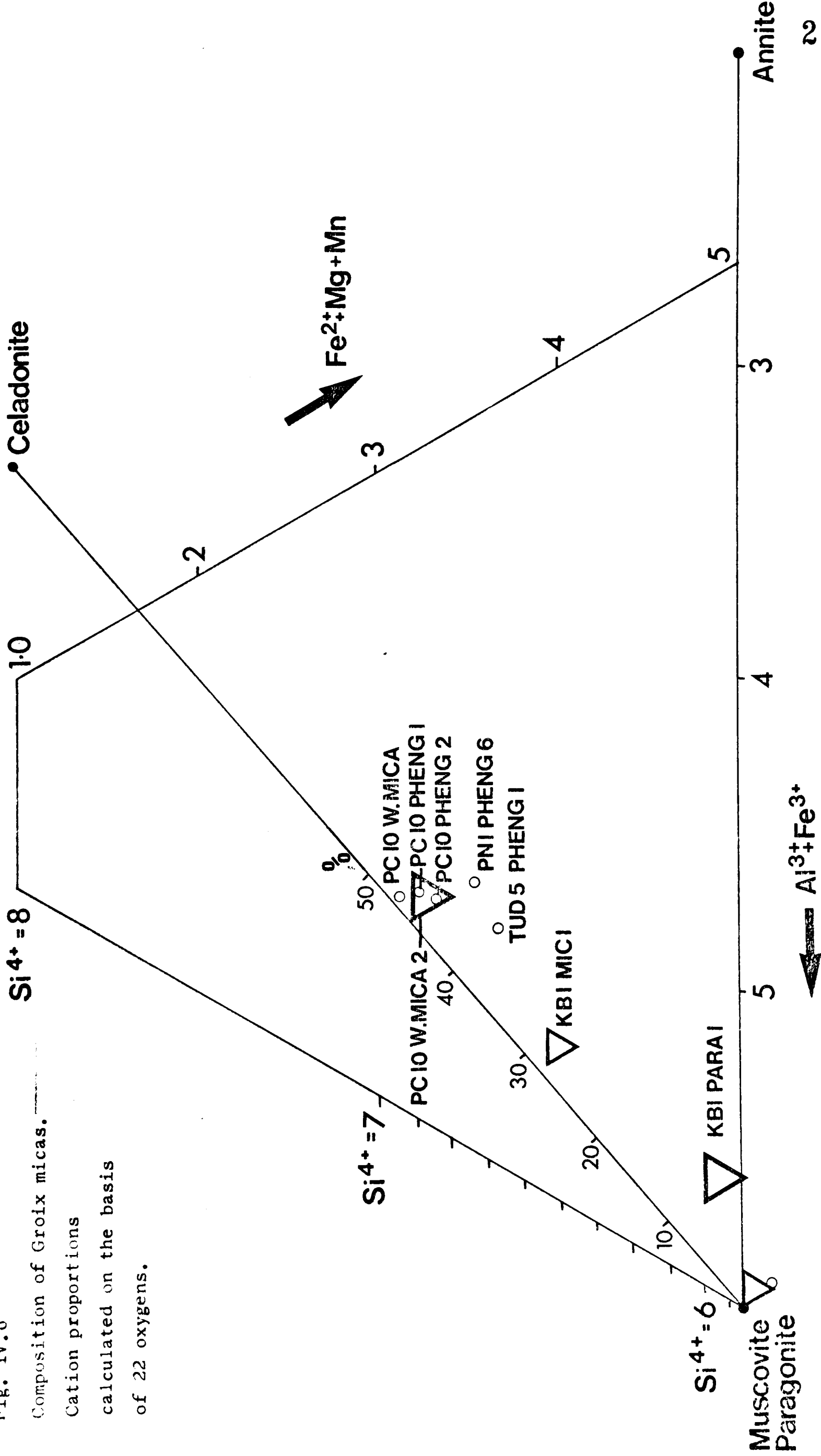
Fig. IV.6

Composition of Groix micas.

Cation proportions

calculated on the basis

of 22 oxygens.



REFERENCES

- ADAMS, C.J.D. A geochronological and related isotopic study of rocks from North-western France and the Channel Islands (United Kingdom). Unpublished D.Phil. Thesis, Univ. of Oxford (1967).
- AUVRAY, B. Relations entre plutonisme acide et volcanisme ignimbritique: exemple des manifestations magmatiques cambriennes du Nord de Bretagne. *Petrologie*, 1, 125-138 (1975).
- BANNO, S. Classification of eclogites in terms of physical conditions of their origin. *Phys. Earth Planet. Inter.*, 3, 405-421 (1970).
- BINNS, R.A. Barroisite bearing eclogite from Nantsdal, Sogn og Fjordane, Norway. *J. Petrology*, 8, 349-371 (1967).
- BLACK, P.M. Mineralogy of New Caledonian metamorphic rocks. 1. Garnets from the Ouegon district. *Contr. Mineral. and Petrol.*, 38, 221-235 (1973).
- BLACK, P.M. Oxygen isotope study of metamorphic rocks from the Ouegon district, New Caledonia. *Contr. Mineral. Petrol.*, 47, 197-206 (1974).
- BLOXAM, T.W. and ALLEN, J.B. Glaucophane-schist, eclogite and associated rocks from Knockormal in the Girvan-Balantrae Complex, South Ayrshire. *Trans. Roy. Soc. Edinburgh*, 64, 1-27 (1958).
- BOETTCHER, A.L. and WYLLIE, P.J. The calcite-aragonite transition measured in the system $\text{CaO-CO}_2\text{-H}_2\text{O}$. *J. Geol.*, 76, 314-330 (1968a).
- BOETTCHER, A.L. and WYLLIE, P.J. Jadeite stability measured in the presence of silicate liquids in the system $\text{NaAlSiO}_4\text{-SiO}_2\text{-H}_2\text{O}$. *Geochim. Cosmochim. Acta*, 32, 999-1012 (1968b).
- BOTTINGA, Y. and JAVOY, M. Oxygen isotope partitioning among the minerals in igneous and metamorphic rocks. *Rev. Geophys. Space Phys.*, 13, 401-418, (1975).
- BRIÈRE, Y. Les Eclogites francaises - Leur Composition mineralogique et chimique; leur origine. *Bull. Soc. Franc. Miner.*, 43, 72-222 (1920).

- BROOKS, C., HART, S.R., HOFMANN, A. and JAMES, D.E. Rb-Sr mantle isochrons from oceanic regions. *Earth and Planet. Sci. Lett.*, 32, 51-61 (1976).
- BROWN, E.H. The Greenschist facies in part of Eastern Otago, New Zealand. *Contr. Mineral. Petrol.*, 14, 259-292 (1967).
- BURNHAM, C.W., HOLLOWAY, J.R. and DAVIS, N.F. Thermodynamic properties of water to 1000°C and 10,000 bars. *Geol. Soc. Amer. Spec. Paper* 132, (1969).
- CARPENTER, M.S.N. and CIVETTA, L. Hercynian high pressure/low temperature metamorphism in the Ile de Groix Blueschists. *Nature*, 262, 276-277 (1976).
- CARTER, S.R. Petrology and geochemistry of the lavas of Mt. Etna, Sicily. Unpublished D.Phil. Thesis, Univ. of Oxford (1976).
- CHAPMAN, H.J. Strontium isotope ratios from the Troodos Ophiolite Complex, Cyprus. Unpublished M.Sc. dissertation, Univ. of Oxford (1974).
- CHINNER, G.A. Chloritoid and the isochemical character of Barrow's zones. *J. Petrology*, 8, 268-282 (1967).
- CLARK, J. and PAPIKE, J.J. Eclogitic pyroxenes, ordered with P2 symmetry. *Science*, 154, 1003-1004 (1966).
- COLEMAN, R.G. and LEE, D.E. Glaucophane-bearing metamorphic rock types of the Cazadero area, California. *J. Petrology*, 4, 260-301 (1963).
- COLEMAN, R.G., LEE, D.E., BEATTY, L.B. and BRANNOCK, W.W. Eclogites and eclogites: their differences and similarities. *Geol. Soc. Amer. Bull.*, 76, 483-508 (1965).
- COGNÉ, J. Schistes cristallins et granites en Bretagne meridionale: Le Domaine de l'Anticlinal de Cornouaille. (These, Strasbourg, 1957). *Mem. expl. Carte geol. de France* (1960).
- COGNÉ, J., JEANNETTE, D. and RUHLAND, M. L'Ile de Groix. Etude structurale d'une serie metamorphique a glaucophane en Bretagne meridionale. *Bull. Serv. Carte geol. Als. Lorr.*, 19, 41-96 (1966).

- COX, A. and DALRYMPLE, G.B. Statistical analysis of geomagnetic reversal data and the precision of potassium-argon dating. *J. geophys. Res.*, 72, 2603-2614 (1967).
- DEER, W.A., HOWIE, R.A. and ZUSSMAN, J. Rock forming minerals. Vol. 3. Sheet silicates. Longmans, London (1962).
- DEER, W.A., HOWIE, R.A. and ZUSSMAN, J. Rock forming minerals. Vol. 2. Chain silicates. Longmans, London (1963).
- ERLANK, A.J. and KABLE, E.J.D. The significance of incompatible elements in Mid-Atlantic ridge basalts from 45°N with particular reference to Zr/Nb. *Contr. Mineral. Petrol.*, 54, 281-291 (1976).
- ERNST, W.G. Stability relations of glaucophane. *Am. J. Sci.*, 259, 735-765 (1961).
- ERNST, W.G. Petrogenesis of glaucophane schists. *J. Petrology*, 4, 1-30 (1963).
- ERNST, W.G. CO₂-poor composition of the fluid attending Franciscan and Sanbagawa low-grade metamorphism. *Geochim. Cosmochim. Acta*, 36, 497-504 (1972).
- ESSENE, E.J. and FYFE, W.S. Omphacite in Californian Metamorphic Rocks. *Contr. Mineral. Petrol.*, 15, 1-23 (1967).
- FAURE, G. and POWELL, J.L. Strontium isotope geology. Springer, New York (1972).
- FELIX, C. Etude petrographique des roches basiques de l'Ile de Groix (Bretagne meridionale). *Ann. Soc. Geol. Belgique*, 92, 359-370, (1969).
- FELIX, C. Etude structuro-mineralogique des pseudomorphes de presmee lawsonite des glaucophanoschistes de l'Ile de Groix (Bretagne - France) *Ann. Soc. Geol. Belgique*, 95, 345-391 (1972).
- FELIX, C. and FRANSDET, A.M. Pseudomorphes a epidote s.l. paragonite, muscovite s.l., chlorite, albite ... de porphyroblastes de lawsonite

(?) dans les glaucophanoschistes de Groix. (Bretagne - France).

Ann. Soc. Geol. Belgique, 95, 323-334 (1972).

FLANAGAN, F.J. 1972 values for international geochemical reference samples.

Geochim. Cosmochim. Acta, 37, 1189-1200 (1973).

GANGULY, J. Activity-composition relation of jadeite in omphacite pyroxene.

Earth Planet. Sci. Lett., 19, 145-153 (1973).

GANGULY, J. and KENNEDY, G.C. The energetics of natural garnet solid

solution. 1. Mixing of the aluminosilicate end-members. Contr.

Mineral. Petrol., 48, 137-148 (1974).

GREEN, D.H. and RINGWOOD, A.E. The genesis of basaltic magmas. Contr.

Mineral. Petrol., 15, 103-190 (1967).

HAYS, J.F. and BELL, P.M. Albite-Jadeite-Quartz equilibrium: A hydrostatic

determination. Carn. Inst. Yrbook, 72, 706-708 (1973).

HEINRICH, K.F.J. X-ray absorption uncertainty. In the electron microprobe.

New York: John Wiley and Sons, 296-377 (1966).

HENSEN, B.J., SCHMID, R. and WOOD, B.J. Activity composition relations in

grossular-pyroxene solid solutions. Contr. Mineral. Petrol., 51,

161-166 (1975).

HOFFMANN, C. Natural and synthetic ferroglaucofan. Contr. Mineral. Petrol.

34, 135-149 (1972).

HOSCHEK, G. The stability of staurolite and chloritoid and their significance

in metamorphism of pelitic rocks. Contr. Mineral. and Petrol., 22,

208-232 (1969).

JÄGER, E. Die alpine Orogenese im Lichte der radiometrischen Altersbesti-

mmung. Eclog. geol. Helv., 66, 11-21 (1973).

JOHANNES, W., BELL, P.M., MAO, H.K., BOETTCHER, A.L., CHIPMAN, D.W., HAYS,

J.F., NEWTON, R.C. and SEIFERT, F. An interlaboratory comparison of

piston-cylinder pressure calibration using the albite breakdown

reaction. Contr. Mineral. Petrol., 32, 24-38 (1971).

- KERRICK, D. Experiment on the upper stability limit of pyrophyllite at 1.8 kb and 3.9 kb water pressure. *Am. J. Sci.*, 266, 204-214 (1966).
- KIENAST, J.R. and TRIBOULET, C. Le chloritoïde dans les paragenèses à glaucophane, albite ou paragonite. *Bull. Soc. fr. Mineral. Cristallogr* 95, 565-573 (1972).
- LIU, J.G. Synthesis and stability relations of epidote $\text{Ca}_2\text{Al}_2\text{FeSi}_3\text{O}_{12}(\text{OH})$. *J. Petrology*, 14, 381-413 (1973).
- MAKANJUOLA, Z.A. The mineralogy and geochemistry of the glaucophane and chloritoid bearing schists and associated rocks of the Ile de Groix, Brittany, France. Unpublished Ph.D. thesis, Univ. of London (1967).
- MAKANJUOLA, Z.A. and HOWIE, R.A. The mineralogy of the glaucophane schists and associated rocks from the Ile de Groix, Brittany, France. *Contr. Mineral. Petrol.*, 35, 83-118 (1972).
- MICHOT, J. and DEUTSCH, S. U/Pb zircon ages and polycyclism of the Gneiss de Brest and the adjacent formations (Brittany). *Eclog. geol. Helv.*, 63/1, 215-227 (1970).
- MIYASHIRO, A. Metamorphism and metamorphic belts. George, Allen and Unwin, London (1973).
- MORGAN, B.A. Petrology and mineralogy of eclogite and garnet amphibolite from Puerto Cabello, Venezuela. *J. Petrology*, 11, 101-145 (1970).
- NAKAMURA, N. Determination of REE, Ba, Fe, Mg, Na and K in carbonaceous and ordinary chondrites. *Geochim. Cosmochim. Acta*, 38, 757-775 (1974).
- NEWTON, R.C. and SMITH, J.V. Investigations concerning the breakdown of albite at depth in the Earth. *J. Geol.*, 75, 268-286 (1967).
- NITSCH, K-H. Das P-T- X_{CO_2} -stabilitätsfeld von Lawsonit. *Contr. Mineral. Petrol.*, 34, 116-134 (1972).
- NORRIS, R.J. and HENLEY, R.W. Dewatering of a metamorphic pile. *Geology*, 4, 333-336 (1976).

- OKA, Y. and MATSUMOTO, T. Study on the compositional dependance of the apparent partition coefficient of iron and magnesium between co-existing garnet and clinopyroxene solid solution. *Contr. Miner. Petrol.*, 48, 115-121 (1974).
- OXBURGH, E.R. and TURCOTTE, D.L. Thermal gradients and regional metamorphism in overthrust terrains with special reference to the Eastern Alps. *Schweiz. miner. petrogr. Mitt.*, 54, 641-662 (1974).
- PANKHURST, R.J. Strontium isotope and geochronological studies of the basic igneous province of N.E. Scotland. Unpublished D.Phil. thesis, Univ. of Oxford (1968).
- PANKHURST, R.J. and O'NIONS, R.K. Determination of Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of some standard rocks and evaluation of X-ray fluorescence spectrometry in Rb-Sr geochemistry. *Chem. Geol.*, 12, 127-136 (1973).
- PEARCE, J.A. and CANN, J.R. Tectonic setting of basic volcanic rocks determined using trace element analyses. *Earth Planet. Sci. Lett.*, 19, 290-300 (1973).
- POPP, R.K. and GILBERT, M.C. Stability of Acmite-Jadeite pyroxenes at low pressure. *Amer. Mineral.*, 57, 1210-1231 (1972).
- POWELL, R. Mineral equilibria in the Leven schists near Fort William, Inverness-shire. Unpublished D.Phil. thesis, Univ. of Oxford (1973).
- POWELL, M. and POWELL, R. An olivine-clinopyroxene geothermometer. *Contr. Mineral. Petrol.*, 48, 249-263 (1974).
- RÅHEIM, A. and GREEN, D.H. Experimental determination of the temperature and pressure dependance of the Fe-Mg partition coefficient for co-existing garnet and clinopyroxene. *Contr. Mineral. Petrol.*, 48, 179-203 (1974).
- RÅHEIM, A. and GREEN, D.H. P, T paths of natural eclogites during metamorphism - a record of subduction. *Lithos*, 8, 317-328 (1975).
- ROBIE, A.A. and WALDBAUM, D.R. Thermodynamic properties of minerals and related substances at 298.15⁰K (25.0⁰C) and one atmosphere (1.013 bars) pressure and at higher temperatures. *Bull. U.S. Geol. Surv.*, 1259 (1968).

- RUHLAND, M. and JEANNETTE, D. La Baie de Ker Bethanie: secteur clef de la structure geologique de l'Ile de Groix. Bull. Serv. Carte geol. Als. Lorr., 19, 97-104 (1966).
- RYBURN, R.J., RAHEIM, A. and GREEN, D.H. Determination of the P,T paths of natural eclogites during metamorphism - record of subduction. A correction to a paper by Raheim and Green (1975). Lithos, 9, 161-164 (1976).
- SCHUILLING, R.D. and WINK, B.W. Stability relations of some titanium minerals (sphene, perovskite, rutile, anatase). Geochim. Cosmochim. Acta, 31, 2399-2411 (1967).
- SPOONER, E.T.C. and FYFE, W.S. Sub-sea floor metamorphism, heat and mass transfer. Contr. Mineral. Petrol., 42, 287-304 (1973).
- SPRY, A. Metamorphic textures. Pergamon Press, (1969).
- SUPPE, J. and ARMSTRONG, R.L. K/Ar dating of Franciscan metamorphic rocks. Am. J. Sci., 272, 217-233 (1972).
- SWEATMAN, J.R. and LONG, J.V.P. Quantitative electron probe microanalysis of rock forming minerals. J. Petrology, 10, 332-379 (1969).
- TAYLOR, H.P. and COLEMAN, R.G. O^{18}/O^{16} ratios of co-existing minerals in glaucophane bearing metamorphic rocks. Geol. Soc. Amer. Bull., 79, 1727-1756 (1968).
- TRIBOULET, C. Mise en evidence de stilpnomelane, biotite verte et phengite chromifere a l'Ile de Groix, Morbihan (France). C.R. Acad. Sci., Paris, 292(D), 1477-1480 (1971).
- TRIBOULET, C. Les glaucophanites et roches associes de l'Ile de Groix (Morbihan, France): etude mineralogique et petrogenetique. Contr. Mineral. Petrol., 35, 83-118 (1974).
- VELDE, B. Phengite micas: synthesis, stability and natural occurrence. Am. J. Sci., 263, 886-913 (1965).
- VELDE, B. Si^{4+} content of natural phengites. Contr. Mineral. Petrol., 14, 250-258 (1967a).

- VELDE, B. Note sure un micaschiste a chloritoide et glaucophane de l'Ile de Groix, Morbihan. Bull. Soc. fr. Mineral. Cristallogr., 90, 265-266 (1967b).
- VELDE, B. Les eclogites de la region nantaise (de Campbon au Cellier, Loire-Atlantique). Bull. Soc. fr. Mineral. Cristallogr., 93, 370-385 (1970).
- VELDE, B. and KOMPROBST, J. Stabilite des silicates d'alumine hydrates. Contr. Mineral. Petrol., 21, 63-74 (1969).
- VERNON, R.H. Metamorphic processes: reactions and microstructure development. Murby, London (1976).
- VIDAL, P. L'axe granitique de Moelan-Lanvaux (sud du Massif Armoricaïn): Mise en evidence, par la methode Rb-Sr, de trois episodes de plutonisme pendant le Palaeozoic inferieur. Bull. Soc. geol. mineral. Bretagne, (C) 4,2, 75-89 (1972).
- VIDAL, P. Premiers donees geochronologiques sur les granites hercyniens du Sud du Massif Armoricaïn. Bull. Soc. geol. Fr., 7, (15), 239-245 (1973).
- VIDAL, P., AUVRAY, B., COGNE, J., HAMEURT, J. and JEANNETTE, D. Donees geochronologique sur la serie spilitique d'Erquy: problemes nouveaux a propos du Brioverian de Bretagne septentrionale. C.R. Acad. Sci. Paris, 273(D), 132-135 (1971).
- VIDAL, P., DEUTSCH, S., MARTINEAU, F. and COGNE, J. Nouvelles donees radiometriques en Baie de Saint-Brieuc. Le probleme d'un socle ante Cadomien nord-armoricaine. C.R. Acad. Sci., Paris, 279(D), 631-634 (1974).
- VOGEL, D.E. and GARLICK, G.D. Oxygen isotope ratios in metamorphic eclogites. Contr. Mineral. Petrol., 28, 183-191 (1970).
- WATERS, D.J. Structural, metamorphic and geochronological studies in the S.E. Tauern. Unpublished D.Phil. Thesis, Univ. of Oxford (1976).
- WINKLER, H.G.F. Petrogenesis of metamorphic rocks. Springer-Verlag, Berlin (1967).

WIT, M.J. de. Metamorphic textures and deformation: A new mechanism for the development of syntectonic porphyroblasts and its implications for interpreting timing relationships in metamorphic rocks. *Geological J.*, 11, 71-100 (1976).

ZEN, E-an. Construction of pressure-temperature diagrams for multicomponent systems after the method of Schreinemakers - a geometrical approach. *U.S. Geol. Surv. Bull.*, 1225, (1966).

ADDENDUM

GUGGENHEIM, E.A. *Mixtures*

Clarendon Press, Oxford (1952).

HARVEY, P.K., TAYLOR, D.M., HENDRY, R.D., and BANCROFT, F.

An accurate fusion method for the analysis of rocks and chemically related materials by X-Ray fluorescence Spectrometry.

X-Ray Spectrom., 2, 33-44 (1973).