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**The Geology of the Hypersthene Gabbro  
of Ardnamurchan Point  
and implications for its evolution  
as an upper crustal  
basic magma chamber**

**VOLUME 1.**

**Simon John Day  
B.A.(Oxon.), FGS**

**Department of Geological Sciences  
University of Durham**

**Thesis submitted for the degree of  
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University of Durham**

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## ABSTRACT

### The Geology of the Hypersthene Gabbro of Ardnamurchan Point and Implications for its evolution as an Upper Crustal Basic Magma Chamber.

S.J.Day.

1989

The Hypersthene Gabbro of Ardnamurchan Point is formed by an outer Marginal Border Group ( MBG ) and a younger Inner Series ( IS ). The MBG is a single large intrusion which corresponds to a high - melt - percentage magma chamber. The IS is dominated by numerous gabbro-norite sheet intrusions which formed a large low - melt - percentage magma body.

The country rocks around the MBG show polyphase metamorphism. An early ( M1 ) phase of high grade metamorphism was followed by sudden cooling and then by hydrothermal metamorphism ( M2 ), related in part to the emplacement of the IS. The sudden cooling was caused by self - propagation of tensile fracture networks containing vigorously convecting hydrothermal fluids. The fracture networks were initiated by tectonic fracturing. The fractures networks also propagated into the MBG and partly preserved the magma chamber boundary layer formed during M1.

The contact of the MBG was approximately stationary during M1. Wall - rock melting occurred in an episodic process triggered by movement on concentric inward - dipping normal faults due to fluctuations in magma pressure. The heat flux  $Q_m$  in the boundary layer was approximately equal to the heat flux  $Q_c$  in the adjacent wall rocks. The preservation of the end - M1 instantaneous metamorphic thermal gradient in the country rocks by the subsequent sudden cooling allows direct measurement of  $Q_c$  and hence of  $Q_m$  (  $8 - 40 \text{Wm}^{-2}$  ) and other parameters of the boundary layer of the MBG magma chamber. The interior of the MBG magma chamber was probably just stably stratified but cooling at the chamber walls produced density currents and slow mixing between the layers. The chamber was not well - mixed: variations in previous crustal contamination of the magmas have been preserved.

The IS shows evidence for interstitial melt expulsion related to the formation of igneous lamination. Hydrothermal circulation in the IS, at up to  $1000^\circ\text{C}$ , produced oxidation of the rocks and may have led to the formation of hydrous melts.

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## VOLUME I

### CHAPTER 1.

A REVIEW OF PREVIOUS WORK ON THE HYPERSTHENE GABBRO OF ARDNAMURCHAN POINT AND ON OTHER LARGE MAFIC INTRUSIONS	p.1
1.1 The Geological and Scientific setting of this research	p.1
1.1.1 An Introduction to the geological setting of the Hypersthene Gabbro of Ardnamurchan Point	p.1
1.1.2 A brief introduction to the whys and wherefores of Magma Chamber Research	p.4
1.1.3 Types of research into magma chambers and the results obtained by each	p.5
1.1.3.1 Geophysical, geodetic and structural studies of active magma chambers, principally those beneath large active volcanoes and rift zones	p.6
1.1.3.2 Petrological studies of quenched materials tapped from magma chambers	p.7
1.1.3.3 Field, petrographic, mineralogical and geochemical studies of large intrusions	p.7
1.1.3.4 The application of physics to magma chamber studies	p.10
1.2 Previous Research on the Hypersthene Gabbro of Ardnamurchan Point	p.13
1.2.1 Early Research and the work of the Geological Survey	p.13
1.2.2 The Complete Re-Investigation of the Hypersthene Gabbro by M.K.Wells	p.19
1.2.3 Further re-interpretations of the structure of the Hypersthene Gabbro and the application of some new techniques	p.22
1.2.4 Advantages of the Hypersthene Gabbro as a "natural laboratory" for the study of subvolcanic magma chambers	p.23

## CHAPTER 2.

THE PRE-HYPERSTHENE GABBRO GEOLOGY OF ARDNAMURCHAN: THE DISTRIBUTION AND COMPOSITION OF POTENTIAL CONTAMINANTS, THE PREVIOUS STRUCTURE OF THE AREA AND STRUCTURAL MARKERS	p.27
2.1 The Lewisian Complex	p.28
2.2 Moine Group Metasediments	p.32
2.3 Mesozoic Sediments	p.37
2.3.1 Triassic and Lower Jurassic	p.40
2.3.2 Middle Jurassic	p.44
2.3.3 Summary of the Mesozoic Sequence and its geochemical characteristics	p.45
2.4 Pre-Hypersthene Gabbro Tertiary Rocks	p.47
2.4.1 The Plateau Basalt Sequence	p.47
2.4.2 Volcaniclastic Rocks	p.56
2.4.3 Early Tertiary basic to felsic intrusions	p.58
2.5 The Pre-Hypersthene Gabbro Structure of Western Ardnamurchan	p.60

## CHAPTER 3.

FIELD AND PETROGRAPHIC EVIDENCE FOR A REVISED SUBDIVISION OF THE HYPERSTHENE GABBRO AND FOR THE AGE RELATIONSHIPS OF ROCKS, METAMORPHISM AND STRUCTURES IN AND AROUND THE PLUTON	p.66
3.1 An Overview	p.66
3.2 The Marginal Border Group (MBG)	p.68
3.2.1 Outcrop distribution and characteristics	p.68

3.2.2	Glendrian Bay and Hill 90	p.68
3.2.2.1	The outer contact of the Hypersthene Gabbro and the rocks of the Marginal Border Group	p.68
3.2.2.2	The country rocks in the Glendrian Bay area, and Minor Intrusions associated with the Pluton	p.85
3.2.2.3	Summary of Age Relationships in the Glendrian Bay area	p.94
3.2.3	Duin Bhain to Sanna Point	p.97
3.2.3.1	M1 rocks and structures in the Contact Zone of the MBG between Duin Bhain and Sanna Point	p.97
3.2.3.2	Country rocks adjacent to the Hypersthene Gabbro between Sanna Point and Duin Bhain	p.105
3.2.3.2	Cone Sheets and Felsite Intrusions between Sanna Point and Duin Bhain	p.106
3.2.3.4	Deformation in the Sanna Point - Duin Bhain area and a summary of the overall age relationships	p.111
3.2.4	Ardnamurchan Point to An Acairseid	p.112
3.2.4.1	The geometry of the western margin of the MBG	p.112
3.2.4.2	Contact Zone Rocks along the western margin of the MBG	p.115
3.2.4.3	Country rocks and minor intrusions on Druim na Cloise	p.122
3.2.4.4	Age Relationships of the MBG and associated rocks between Ardnamurchan Point and Gharblach Mhor	p.123
3.2.5	An Acairseid to Kilchoan	p.125
3.2.5.1	The Orientation of the Contact and the effects of Radial and Concentric faults upon it	p.125

3.2.5.2	Contact Zone Rocks along the Southern Margin of the Hypersthene Gabbro	p.138
3.2.5.3	High degree partial melting of country rocks along the southern margin of the Marginal Border Group	p.147
3.2.5.4	The Structure of the Country Rocks away from the Contact in the An Acairseid - Kilchoan sector, and the Minor Intrusions which cut them	p.151
3.2.5.5	Metamorphism on the southern side of the Hypersthene Gabbro and its Age Relationships to Deformation and the Emplacement of Minor Intrusions	p.166
3.2.5.6	The Problems of Spatially Limited and Temporally Extended Periods of Activity in the determination of Age Relationships on the southern margin of the Hypersthene Gabbro	p.181
3.2.6	Glebe Hill and the Glas Eilean fault	p.182
3.2.7	A summary of the rocks and age relationships, up to the middle of M2, observed in the Contact Zone of the Marginal Border Group and the adjacent country rocks	p.200
3.3	The geometry and origin of the contact between the Marginal Border Group and the Inner Series	p.205
3.3.1	Field, petrographic and mineralogical criteria for distinguishing rocks of the Marginal Border Group from those of the Inner Series	p.205
3.3.2	Contact relationships between Marginal Border Group and Inner Series Rocks	p.208
3.4	The Inner Series	p.218
3.4.1	Gabbronorites : cumulate sequence or intrusive sheet complex ?	p.218

3.4.2	Pre-gabbro-norite layered and laminated rocks	p.233
3.4.2.1	Sanna Point and Sanna Bay	p.235
3.4.2.2	Pre-gabbro-norite rocks at An Acairseid	p.245
3.4.2.3	Opaque-Rich and Laminated rocks on and to the south of Beinn na Seilg	p.248
3.4.2.4	Other occurrences of pre-gabbro-norite rocks in the Inner Series, and some comments on their overall distribution	p.250
3.4.3	Secondary or Replacive Olivine Pyroxenites	p.253
3.4.4	Blastoporphyritic granular dolerite and microgabbro-norite sheet intrusions	p.263
3.4.5	Post - metamorphic minor intrusions	p.267
3.4.6	Metamorphism of the Inner Series rocks	p.268
3.4.7	Deformation of the Inner Series and criticisms of the central subsidence model for layering orientation	p.279
3.4.8	A summary of the geology of the Inner Series and its age relationships to adjacent rocks	p.286
3.5	A summary of the geological history of the Hypersthene Gabbro	p.291

## VOLUME II

### CHAPTER 4.

MAGMATIC PROCESSES IN THE MARGINAL BORDER GROUP MAGMA CHAMBER DURING M1 METAMORPHISM	p.295
4.1 Theoretical models of magma chamber boundary layers and their application to the Marginal Border Group	p.295
4.1.1 Introduction, and the concept of a 'rigidus'	p.295
4.1.2 Types of Magma Chamber Boundary Layer	p.298
4.1.3 The limitations of theoretical calculations regarding the behaviour of magma chamber boundary layers, and an outline of an alternative approach to the study of them	p.305
4.2 The Felsic Rocks of the Marginal Border Group : Geochemistry, Petrogenesis and Physical Properties at the Time of Emplacement	p.309
4.2.1 Overall Distribution of the Felsic Rocks	p.309
4.2.2 M1 microgranodiorites and M2 felsites	p.311
4.2.2.1 Geochemistry and mineralogy	p.311
4.2.2.2 Mineralogy and textures of the porphyrocrysts in the M2 felsites	p.335
4.2.2.3 Summary of evidence for the origin of the subalkaline granitoids	p.347
4.2.3 Compositionally Anomalous Granitoid Rocks and Country Rock Xenoliths	p.348
4.2.3.1 The Main Types of Anomalous Granitoid Rocks within the Contact Zone	p.349
4.2.3.2 Country Rock Xenoliths	p.350
4.2.3.3 Geochemistry and Petrogenesis of Anomalous Granitoids and Country Rock Xenoliths - 1 : General Considerations	p.354

4.2.3.4	Geochemistry and Petrogenesis of Anomalous Granitoids and Country Rock Xenoliths - 2 : Petrogenesis of each of the Granitoid Types and of the Xenoliths	p.362
4.2.3.5	Conclusions : Implications of the petrogenesis of the Anomalous Granitoids and Xenoliths for processes at the wall of the MBG magma chamber	p.368
4.2.4	Rheomorphic Breccias	p.369
4.2.4.1	Composition of the Rheomorphic Breccias	p.369
4.2.4.2	Development of Textures and Structures in the Rheomorphic Breccias	p.379
4.2.5	Estimates of the Conditions of Melting during the formation of the anomalous microgranitoids, M1 and M2 subalkaline microgranitoids and Rheomorphic Breccias	p.385
4.2.5.1	Temperature and pressure constraints from the quartz - tridymite inversion	p.386
4.2.5.2	Pressure - temperature constraints from phase assemblages in the subalkaline granitoids	p.386
4.2.5.3	Plagioclase - Melt geothermometry	p.387
4.2.5.4	Geological Constraints on the lower value of P(total)	p.390
4.2.5.5	Synthesis of the constraints on the Temperature and Pressure of formation of the MBG granitoids	p.393
4.2.6	Implications of the petrogenesis of the felsic rocks for the behaviour of the MBG magma chamber at the end of M1	p.395

4.3	Homogenous Basic and Intermediate Rocks of the Marginal Border Group	p.403
4.3.1	General Geochemistry of the Basic and Intermediate Rocks	p.403
4.3.2	The Nature of Crustal Contamination in the Homogenous Marginal Border Group Rocks : Contamination Within or Beneath the MBG Magma Chamber ?	p.414
4.3.2.1	Assimilation before, during or after fractional crystallisation ?	p.414
4.3.2.2	Identification of the Contaminant Lithologies	p.415
4.3.2.3	Implications of the Crustal Contamination for Convection in the MBG Magma Chamber	p.428
4.3.3	Temperatures of the basic and intermediate magmas within the MBG magma chamber during M1 and estimated density differences between the basic, intermediate and subalkaline granitoid magmas	p.433
4.3.4	Relationships of the MBG basic rocks to the later M2 minor intrusions	p.438
4.3.5	Conclusions on the Petrogenesis of the MBG Homogenous Basic and Intermediate rocks and Implications for the behaviour of the MBG Magma Chamber	p.441
4.4	Heterogenous Hybrid Rocks in the Marginal Border Group	p.445
4.4.1	Geochemistry of the Heterogenous Hybrid Rocks	p.445
4.4.1.1	Compositions of and Compositional Variation in the End - Member Magmas	p.445
4.4.1.2	Large - scale magma movements associated with the formation of the heterogenous hybrid rocks	p.453

4.4.2	Mechanisms of the formation of the hybrid magmas and implications for variations in the thermal regime at the margin of the intrusion	p.459
4.5	Summary : The behaviour of the Marginal Border Group Magma Chamber during the later part of M1	p.467
4.5.1	Petrogenesis of the M1 - age components of the Marginal Border Group and associated minor intrusions and implications of each for the behaviour of the magma chamber	p.467
4.5.1.1	The Anatectic Rocks and Xenoliths	p.467
4.5.1.2	Homogenous Basic to Intermediate Rocks	p.470
4.5.1.3	Heterogenous hybrid rocks	p.471
4.5.2	A model for the magma circulation within the MBG magma chamber and the effects of episodic wall rock melting	p.471
4.5.2.1	Magma fluxes within the MBG magma chamber	p.471
4.5.2.2	Effect of episodic wall - rock melting upon the compositional evolution of the MBG magma chamber	p.475
4.5.2.3	To what extent has the MBG boundary layer been preserved ?	p.480
4.5.3	Implications of the magma circulation pattern proposed in section 4.5.1 for heat flow from the magma chamber and estimates of the temperature difference across the boundary layer	p.481 ✓
4.5.3.1	Implications of compositional variation in boundary layers for boundary layer behaviour	p.481
4.5.3.2	Estimate of a value for the temperature difference across the thermal boundary layer	p.482

CHAPTER 5.

METAMORPHISM ASSOCIATED WITH THE MARGINAL BORDER GROUP AND IMPLICATIONS FOR THE MASS AND HEAT FLUXES IN THE CONTACT AUREOLE AND IN THE BOUNDARY LAYER OF THE MBG MAGMA CHAMBER	P.487
5.1 Conditions of metamorphism during M0	p.488
5.2 Conditions of M1 Metamorphism	p.493
5.2.1 The width of the M1 aureole and variations in the grade of metamorphism within it	p.493
5.2.1.1 The relationship between the present day width of the M1 contact aureole and its width perpendicular to the contact at the end of M1	p.493
5.2.1.2 Metamorphic reactions and apparent isograds in the M1 aureole	p.495
5.2.2 A preliminary estimate of the temperature profile through the contact aureole at the end of M1	p.501
5.2.3 The oxidation state of rocks affected by M1 metamorphism and fluid flow in the M1 aureole	p.505
5.2.4 Thermal modelling of the M1 metamorphic event	p.513
5.2.4.1 An estimate of the heat fluxes $Q_c$ and $Q_m$ at the end of M1, and of the heat transfer coefficient $h$ of the boundary layer	p.513
5.2.4.2 Implications of the value of $Q_m$ for mass fluxes in the boundary layer of the MBG magma chamber at the end of M1	p.516
5.2.4.3 An estimate of the minimum duration of M1	p.522

5.3	The quenching event(s) at the end of M1	p.530
5.3.1	The tensile fracture network in the M1 contact aureole and in the minor felsic intrusions of early M2 age	p.530
5.3.2	Application of the theory of thermal contraction cracking and cracking front propagation to the quenching event	p.534
5.3.2.1	Thermal contraction cracking and self-propagating hydrothermal systems	p.534
5.3.2.2	An alternative method of estimating the cracking front velocity $u$	p.540
5.3.2.3	Concluding remarks on the estimated value of $u$ , and an estimate of the heat flux at the cracking front	p.547
5.3.3	Hydrothermal cooling and the solidification of the Marginal Border Group	p.550
5.3.3.1	A summary of the problems posed by the characteristics of the Marginal Border Group	p.550
5.3.3.2	The propagation velocity $u$ of the hydrothermal system in the Marginal Border Group	p.553
5.3.3.3	Causes of the differences in the textures of the devitrified felsic and holocrystalline mafic rocks of the MBG	
5.3.3.4	Cracking and interstitial melt migration in the Marginal Border Group	p.556
5.3.3.5	Rates of solidification and rates of crystal fractionation during the solidification of the MBG basic rocks	p.560
5.3.3.6	The relationship between the cracking front velocity $u$ and the maximum distance magmas were able to flow down the wall of the MGB magma chamber as it solidified, and implications for the interpretation of the magmatic structure of the MBG	p.562

5.3.3.7	Summary and conclusions	p.565
5.4	M2 Metamorphism in the contact aureole and within the MBG	p.567
5.4.1	Conditions of M2 metamorphism at the peak of metamorphism and towards the end of M2	p.567
5.4.2	Implications of the variation in peak metamorphic temperatures for hydrothermal circulation during M2	p.572
5.5	Possible causes of the changes in heat flow mechanisms during metamorphism in and around the Marginal Border Group	p.574
CHAPTER 6		
	MAGMATIC AND METAMORPHIC PROCESSES IN THE INNER SERIES OF THE HYPERSTHENE GABBRO	p.583
6.1	Differences between petrological and geophysical models of basic magma chambers in the upper crust : does the Inner Series correspond to a 'geophysicist's magma chamber' ?	p.583
6.3	Sub - rigidus and other magmatic processes and their compositional effects on the gabbronorites and related rocks of the inner series	p.585
6.2.1	Other causes of compositional variation in the gabbronorite suite	p.585
6.2.1.1	Were the rocks of the gabbronorite suite derived from cogenetic magmas ?	p.586
6.2.1.2	Cryptic variation in the gabbronorite suite	p.592

6.2.1.3	Compositional variations due to super - rigidus processes in the gabbronorite suite rocks	p.602
6.2.2	Late magmatic ('post - cumulus' or sub - rigidus) processes in the gabbronorite suite rocks	p.610
6.2.2.1	The formation of 'post - cumulus' lamination in the gabbronorites, and associated mineralogical and bulk compositional changes	p.610
6.2.2.2	Pyroxenite formation	p.624
6.3	Compositional relationships of the different groups of rocks within the Inner Series and their relationships to the other rocks of the Hypersthene Gabbro	p.629
6.3.1	Pre - gabbronorite Inner Series rocks	p.629
6.3.1.1	Augite - anorthite / bytownite gabbros or eucrites	p.629
6.3.1.2	Granular magnetite - anorthosites and magnetite - troctolites	p.635
6.3.1.3	Coarse - grained anorthosites	p.635
6.3.1.4	Laminated gabbros and olivine gabbronorite	p.637
6.3.2.1	The Inner Series and the Marginal Border Group	p.637
6.3.2.2	The gabbronorite suite rocks and late - M2 and younger rocks	p.638
6.3.3	Conclusions regarding the magma supply to the Hypersthene Gabbro during its evolution	p.642

6.4	Ultra - High Grade Hydrothermal Metamorphism of the Inner Series rocks	p.643
6.4.1	Two - pyroxene geothermometry of metamorphic assemblages in Inner Series rocks : temperatures of formation of the metamorphic assemblages and of very high grade hydrothermal fluid circulation in the Inner Series	p.646
6.4.1.1	Errors in two - pyroxene geothermometry and the nature of the temperatures obtained	p.646
6.4.1.2	Temperatures obtained from different textural types of pyroxene	p.649
6.4.1.3	The evolution of pyroxene compositions and textures with time during metamorphism of the Inner Series rocks and the ending of diffusive re - equilibration of compositions and textures	p.655
6.4.2	Metamorphic reactions in the Inner Series rocks and implications for $f_{O_2}$ , $P_{H_2O}$ and dissolved cation concentrations in the coexisting fluid phase	p.658
6.4.2.1	The formation of symplectitic intergrowths of orthopyroxene and oxide phases and the oxidation of olivine in the Inner Series rocks	p.658
6.4.2.2	Magnetite reduction	p.662
6.4.2.3	Augite oxidation	p.663
6.4.2.4	Stability relationships of high - temperature hydrous phases in the Inner Series rocks	p.664
6.4.2.5	Metasomatism and the scale of internal buffering of fluid compositions in the Inner Series rocks	p.667
6.4.2.6	Quantitative estimates of $P_{H_2O}$ and $f_{O_2}$ during very high grade hydrothermal alteration of the Inner Series rocks	p.669
6.4.2.7	Conclusions on the mineralogical and compositional effects of hydrothermal	

- fluid circulation at very high  
temperatures through the Inner Series p.673
- 6.4.3 A history of late stage crystallisation and  
high - temperature hydrothermal metamorphism  
in the Inner Series p.674
- 6.5 Conclusions : magmatic processes in a low - melt -  
percentage magma body and their relationship to  
very high temperature hydrothermal circulation p.678
- 6.5.1 The problems of interpreting overall  
compositional trends in the gabbronorite  
suite of the Inner Series p.678
- 6.5.2 A possible relationship between very high  
temperature hydrothermal circulation and sub  
- rigidus feldspar dissolution during the  
formation of lamination and metasomatic  
pyroxenites p.679
- 6.5.3 Controls on the occurrence of very high  
temperature hydrothermal circulation in mafic  
rocks p.681

## CHAPTER 7

- EMPLACEMENT MECHANISMS OF THE VARIOUS COMPONENTS OF THE  
HYPERSTHENE GABBRO AND OF THE ASSOCIATED MINOR  
INTRUSIONS p.683
- 7.1 The overall geometry and initial emplacement of  
the MBG magma chamber p.684
- 7.1.1 Constraints on the geometry of the MBG magma  
chamber above and below the present level of  
exposure p.684
- 7.1.1.1 Geophysical models for the sub -  
surface structure of the Ardnamurchan  
central complex p.684

7.1.1.2	Evaluation of the alternative geophysical models using geochemical data	p.688
7.1.1.3	A summary of geological constraints on the original extent of the MBG magma chamber above the present level of exposure and on its overall geometry	p.690
7.1.2	Initial emplacement of the MBG magma chamber : evidence from early structures	p.691
7.2	Deformation around the MBG magma chamber after its initial emplacement	p.695
7.2.1	M1 and M2 faulting associated with the MBG magma chamber	p.695
7.2.1.1	The concentric inward - dipping normal fault system	p.696
7.2.1.2	Low angle concentric faults	p.699
7.2.1.3	Concentric high angle reverse faults	p.703
7.2.2	The emplacement mechanisms and associated stress fields of cone sheets and other minor intrusions related to the MBG magma chamber	p.703
7.2.2.1	Cone sheets and other concentric minor intrusions	p.703
7.2.2.2	Radial dykes	p.712
7.2.3	Causes of variation in the stress field around the MBG magma chamber during M1 and the early part of M2 and the relationship between magma pressure in the magma chamber and episodic melting at the walls of the chamber during the latter part of M1	p.713

- 7.3 The emplacement of the gabbronorite sheets of the Inner Series p.717
- 7.4 The regional stress field of the British Tertiary Volcanic Province and the asymmetry of the Hypersthene Gabbro p.720
- 7.5 The structural development of the Hypersthene Gabbro p.723

## CHAPTER 8

CONCLUSIONS : INTERRELATIONSHIPS BETWEEN STRUCTURAL ,  
MAGMATIC AND THERMAL ASPECTS OF THE EVOLUTION OF THE  
HYPERSTHENE GABBRO AND THEIR IMPLICATIONS FOR MAGMATIC  
PROCESSES IN OTHER UPPER CRUSTAL MAGMA CHAMBERS

p. 725

- 8.1 The importance of the MBG for theoretical and analogue studies of magma chamber processes p.727
- 8.2 A summary of the evolution of the Hypersthene Gabbro p.729
- 8.3 To what extent are the results of the present study applicable to other upper crustal magma bodies ? p.735
- 8.3.1 Constraints from contact metamorphism , crustal contamination and the compositions of granites in central complexes on values of Q in other basic and ultrabasic plutons p.735

- 8.3.2 The frequency of occurrence of sudden or catastrophic cooling events associated with the development of tensile fracture network - based hydrothermal systems around upper crustal plutons p. 737
- 8.3.3 The importance of episodic melting in the formation of anatectic magmas in upper crustal magma chambers p. 737
- 8.4 Future work p. 740
  - 8.4.1 Further work on the Hypersthene Gabbro p. 740
  - 8.4.2 Implications of the present study for research on magma chamber processes p. 741
- REFERENCES p. 743

Volume 3.

Appendix 1: Analytical methods.	p.777
A1.1 Bulk Rock Analysis Methods: Introduction.	p.777
A1.2 Sample Preparation.	p.778
A1.3 ICP - AES analysis.	p.780
A1.3.1 Analytical procedure.	p.780
A1.3.2 Replicate analyses and analyses of International Reference Materials by ICP - AES.	p.783
A1.4 Spectrophotometric analysis for Phosphorous.	p.785
A1.5 XRF analysis.	p.787
A1.5.1 XRF sample preparation.	p.787
A1.5.2 Analysis conditions.	p.788
A1.5.3 XRF data processing.	p.789
A1.5.3.1 FINAL.	p.789
A1.5.3.2 CBI.	p.791
A1.5.3.3 The nature of the XRF major element data in Appendix 2.	p.792
A1.5.3.4 Trace element analysis using the K - Factor method.	p.793
A1.5.3.5 Comparison of XRF and ICP - AES data.	p.794
A1.6 Instrumental Neutron Activation Analysis.	p.796
A1.7 Electron Microprobe analysis.	p.800
Tables for Appendix 1	p.801

Appendix 2: Bulk Rock Analyses.	p.809
A2.1 Pre - Hypersthene Gabbro Country Rocks.	p.810
A2.2 Marginal Border Group Rocks.	p.828
A2.3 Inner Series Rocks.	p.869
A2.4 Post - Hypersthene Gabbro Rocks.	p.890
A2.5 Instrumental Neutron Activation Analysis Data	p.897
A2.6 Locations and Descriptions of samples analysed by XRF and ICP methods	p.914

Appendix 3:Electron Microprobe Data.	p. 937
A3.1 Marginal Border Group Rocks.	p. 939
A3.2 Inner Series Rocks.	p. 987
A3.3 Miscellaneous Samples.	p.1119

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# 1. A review of previous work on the Hypersthene Gabbro of Ardnamurchan Point and on other Large Mafic Intrusions

## 1.1 The Geological and Scientific setting of this research

### 1.1.1 An introduction to the geological setting of the Hypersthene Gabbro of Ardnamurchan Point.

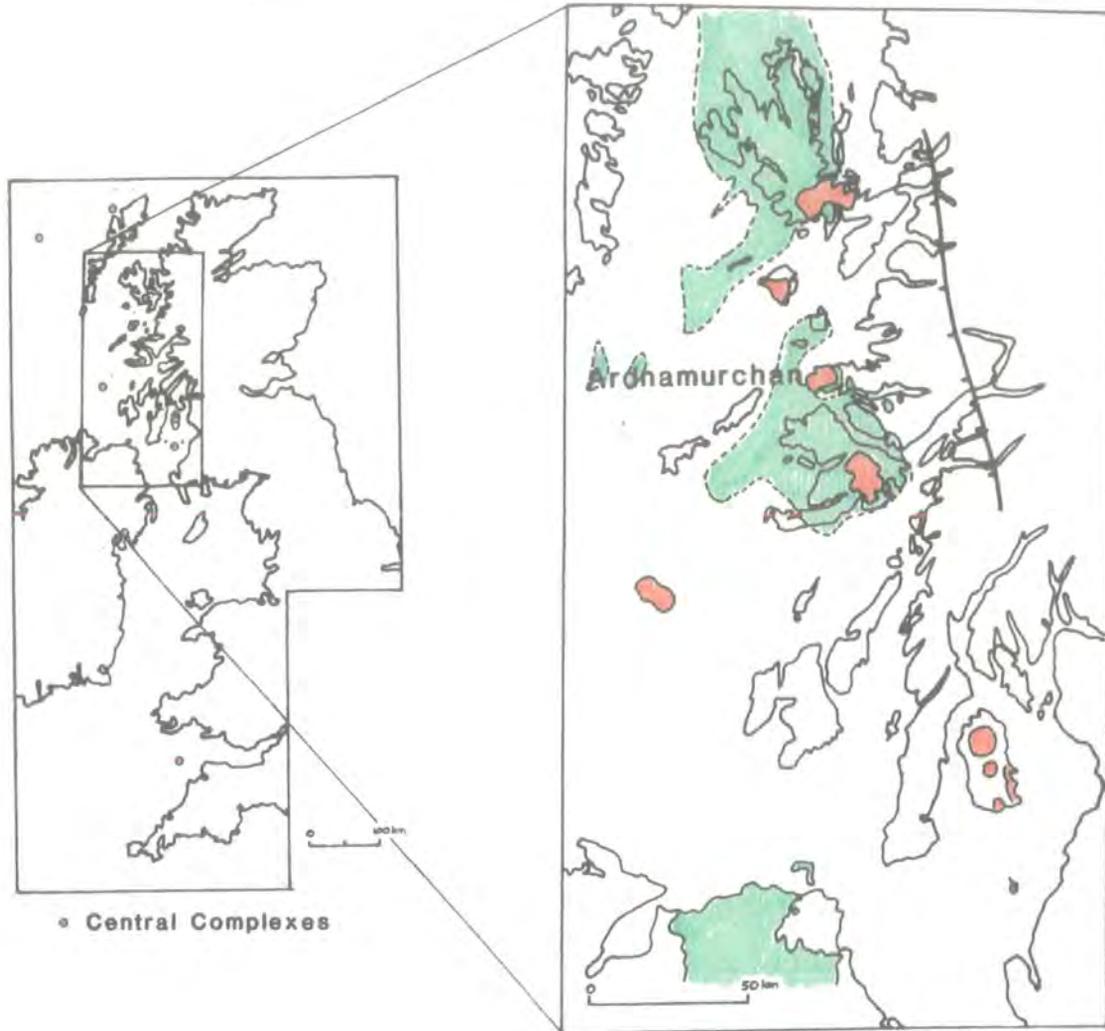
The aim of the work presented in this thesis is to study magmatic and related structural and metamorphic processes involved in the evolution of large subvolcanic mafic intrusions and the magmas and rocks within and passing through them during their active lives as magma chambers.

The means chosen to do this is an integrated study of many aspects of the geology of the Hypersthene Gabbro of Ardnamurchan Point, one such intrusion in the British Tertiary Igneous Province. This pluton, which will usually be referred to as the Hypersthene Gabbro ( note capital letters ) in this work, is in the Tertiary central complex exposed at the western end of the Ardnamurchan peninsula ( see Fig 1.1 for location ). Like most of the other Tertiary igneous centres, which are spaced at intervals of 20 to 100 kilometres through the coastal areas and adjacent seas of northwest Britain, the Ardnamurchan complex is composed of a variety of intrusive and extrusive rocks ( Fig 1.2 ). The intrusive rocks of the centre, consisting of very numerous basic to felsic minor intrusions ( mainly cone sheets ) as well as the volumetrically dominant plutonic rocks, are emplaced partly into pre - Tertiary rocks and partly into Tertiary extrusive rocks. Most of the latter formed the lower parts of the large central volcano which was the surface expression of the centre during its active life. The pre - Tertiary rocks are the late Precambrian psammites and semipelites of the low - grade metamorphic Moine Series and the Triassic and Jurassic sediments which unconformably overlie these, whilst the Tertiary extrusive rocks consist of an older alkaline plateau basalt sequence and younger tholeiitic basalts and various pyroclastic and epiclastic rocks which together form the remnants of the central volcano ( see Fig 1.2 and Chapter 2 ). Geophysical work ( Bott & Tuson 1973, Barrett 1987, Harrison 1987 ) has shown that the gravity and magnetic anomalies associated with the complex correspond to an anomalously dense and magnetic body only 3 to 6 kilometres deep, which probably represents the true thickness of the complex ( see Chapter 7 ). Palaeomagnetic data and isotopic ages reported in Dagley et al. ( 1984 ) suggest that the entire complex formed within the period from 61 to 58.2 Ma.



**Fig. 1.1 Location of the Ardnamurchan Central Igneous Complex within the British Tertiary Igneous Province**

**Based on Richey 1961, with additional information from Binns et al. 1974**



**KEY**



**Central Complex Intrusions**



**Plateau Basalts and Central Complex volcanic rocks**

**Offshore limits of plateau basalts in Sea of the Hebrides from Binns et al. 1974**



**Possible original eastern limit of plateau basalts ( Watson 1985 )**

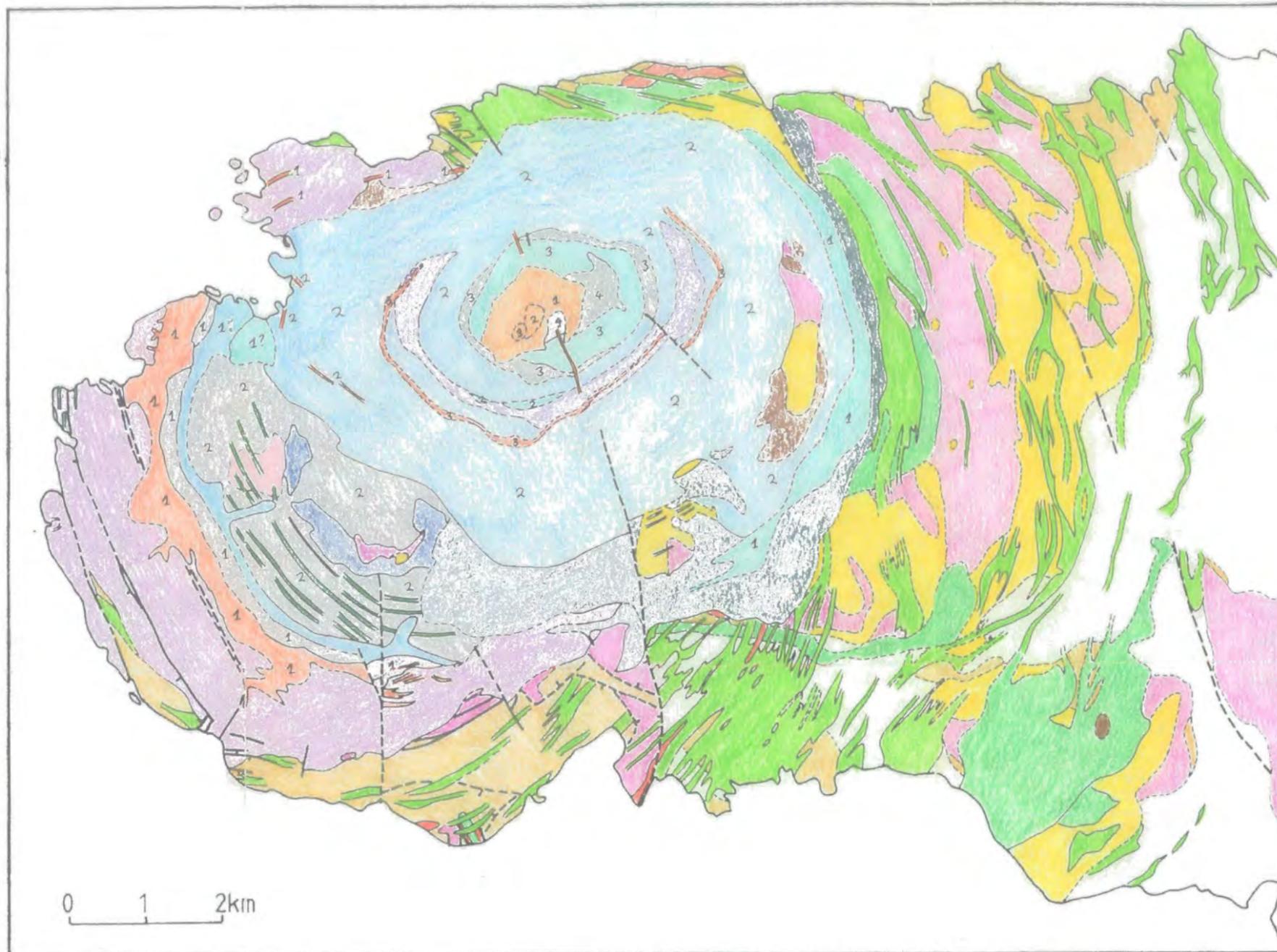
Fig. 1.2 Geological map of the Ardnamurchan Central Complex, W. Scotland ( after Richey et al. 1930, Skelhorn & Elwell 1966 Day 1985, with alterations and additions ).

KEY

-  Late felsite and dolerite dykes
-  Quartz monzonite (3) and tonallite (1 & 2).
-  Quartz Gabbro
-  Laminated Microgabbro
-  Quartz Gabbro.
-  Hybrid granophyre/quartz dolerite.
-  Biotite Eucrite.
-  Laminated Microgabbro.
-  Hybrid granophyre/quartz dolerite.
-  Great Eucrite ( coarse olivine leucogabbro )
-  Laminated Gabbro.
-  Heterogenous quartz- to olivine - gabbro.
-  Small gabbro intrusions of uncertain age.

-  Olivine Leucogabbro
-  Later cone sheets.
-  Grigadale Granophyre.
-  Quartz gabbros and microgabbros.
-  Quartz Gabbro.
-  Sgurr nam Meann hybrid granophyre/microdiorite/quartz dolerite.
-  Old Gabbro of Aodann
-  Hypersthene Gabbro of Ardnamurchan Point.

-  Early cone sheets.



PRE - TERTIARY ROCKS

-  Other early basic to intermediate intrusions.
-  Early microgranitic and felsitic intrusions
-  Pyroclastic rocks: vent agglomerates, ignimbrites, laharic deposits and epiclastic sediments.
-  Basalts, mainly alkaline to transitional plateau basalts.
-  Mesozoic sediments ( Triassic to Middle Jurassic ).
-  Moine Group.
-  Major faults.

Notes: (1) Numerals refer to distinct intrusions of similar rock types. (2) Most dykes omitted. (3) Only principal swarms of cone sheets marked in areas to north and south of the main plutonic complexes: individual sheets are less than 20m thick.

The Hypersthene Gabbro itself occupies an arcuate area at the extreme western end of the peninsula and is the earliest of the large, mainly gabbroic plutons in the complex ( Fig 1.2, which is largely based on the maps of the geological survey workers, Richey et al. 1930 ). It has not received as much attention from geologists as the intrusions of the Rhum and Skye central complexes to the north, but sufficient work had been carried out prior to this study to suggest that a re - examination of the Hypersthene Gabbro in the light of recent theories of magma chamber processes would provide useful insights into the history and behaviour of the ancient magma chamber or chambers which it represented and also provide empirical observations to be used as a basis for testing some of these theories. Before reviewing previous work on the intrusion it is therefore appropriate to consider the reasons for studying magma chambers and also examine the various approaches which have been adopted in these studies and the different types of information which can be extracted from each.

### **1.1.2 A brief introduction to the whys and wherefores of Magma Chamber Research**

Magma chambers may be defined as bodies of wholly and/or partially molten rock in otherwise solid or near - solid surroundings that are at a lower temperature and within which magmas undergo compositional evolution during a protracted period of residence, mainly as a result of loss of heat. They are of interest to the geologist, and to the igneous petrologist in particular, for a number of reasons, of which those most relevant to the present work are:

1) As the definition above ( which is designed to exclude magma conduits ) indicates, magma chambers are sites of magma evolution, by processes such as crystal fractionation, wall - rock assimilation and magma mixing. They are therefore of direct interest to what may be regarded as the central problem of igneous petrology, that of explaining the diversity of igneous rock compositions encountered in nature. In the past, indeed, fractionation processes in magma chambers, and especially crystal fractionation, have been regarded as by far the most important cause of compositional diversity within suites of igneous rocks ( Bowen 1928 ). In more recent years the importance of contamination in magma conduits has been recognised ( Patchett 1981, Thompson et al. 1986, Huppert & Sparks 1985 ) and, perhaps more fundamentally, advances in high - pressure experimental petrology, geochemistry and isotope geology have led to a recognition of the importance of variations in conditions and degrees of primary partial melting ( Klein & Langmuir 1987, McKenzie & Bickle 1988 ) and, perhaps, source rock composition ( White & Chappell 1977 ) in producing variation

within suites of igneous rocks; variation between suites has of course almost always been explained in these ways. Nevertheless, the study of magma chambers has retained an important place in igneous petrology. In part this is because, except for those rocks corresponding to truly primary melts ( and one always has to prove that these have not been affected by fractionation in a magma chamber ), the geochemical effects of magma chamber processes are superimposed upon, and may potentially mimic ( O'Hara & Mathews 1981 ) the effects of earlier processes. An understanding of the physical processes which operate in magma chambers, and of the compositional effects that they produce, is therefore useful in identifying the characteristics of, and variations in, primary melting regimes and source rocks, including contaminants.

2) When present, subvolcanic magma chambers are likely to have a strong influence on the activity of the volcanoes which they feed, both through their influence on the composition and rheology of the magma supply to the volcano above, and therefore on the style of eruptions developed, and through the effect on the structure of the volcano of the stress field produced by the pressure of the magma in the chamber ( Anderson 1936, Gudmundsson 1988, Chevallier & Verwoerd 1988 ). In addition, events in subvolcanic magma chambers such as convective overturns and replenishment with hot, primitive magma may trigger eruptions ( Sigurdsson & Sparks 1981, Turner, Huppert & Sparks 1983 ). Finally, shallow magma chambers are the dominant source of heat for hydrothermal activity in volcanic areas; conversely, hydrothermal circulation is potentially a highly efficient mechanism of heat loss from magma chambers ( Lister 1983a, Strens & Cann 1986 ).

3) The hot and in many cases convecting contents of large magma chambers in the crust are potentially very important sources of heat for crustal melting and metamorphism. Conversely, the processes of heating, metamorphism and melting of the surrounding crustal rocks control magma evolution to a considerable extent, through their effects on rates of heat loss and hence crystallisation, and on contact temperatures and hence rates and mechanisms of crustal contamination ( Irvine 1970, Huppert & Sparks 1988b ).

### 1.1.3 Types of research into magma chambers and the results obtained by each.

With the exception of lava lakes, which may be regarded as subaerial and hence extremely atypical magma chambers ( Helz et al. 1989 ), magma chambers are by nature inaccessible to direct study. A wide range of indirect studies has been carried out which can be divided into four principal categories. In the discussion of these below, attention will be focussed on studies of predominantly basic magma chambers, because it is this type of chamber that is represented by the Hypersthene Gabbro.

#### **1.1.3.1. Geophysical, geodetic and structural studies of active magma chambers, principally those beneath large active volcanoes and rift zones.**

Various combinations of geophysical surveying techniques have been used to locate regions of partial and total melt beneath volcanically active areas, notably in Iceland, Hawaii ( Ryan 1988 ). and on the East Pacific Rise ( Bratt & Solomon 1984, Detrick et al. 1987 ). The most surprising result of these studies, which have the limitation that they only determine the presence and approximate abundance of a melt phase, not its composition, is the discovery that large, high - melt - percentage magma chambers are much less common than crystal mush bodies ( Iyer 1984 ) and swarms of small magma bodies ( Ryan 1987 ), even at shallow depths.

Studies of deformation and structural phenomena associated with subvolcanic magma chambers have been carried out by classical field mapping, high - precision surveying ( e.g. Tryggvason 1986 ) and monitoring of natural seismicity ( e.g. Okaka et al 1981 ). When carried out in 'real time', these studies provide very important information on the timescales of magma chamber replenishment and intrusion. Although the inversion of deformation data in particular to obtain the geometry and associated stress field of a buried magma chamber is a complex procedure that at present requires the use of far too many simplifying assumptions to make any pretence of accuracy ( Chevallier & Verwoerd 1988, Cullen et al. 1987 ), these studies provide a very important link between studies of contemporary volcanic activity and research into the structure of ancient subvolcanic intrusions.

Geophysical studies, of microgravity, electrical conductivity, heat flow and microseismicity ( Foulger 1988 ) are also the principal means, along with borehole studies, of investigating hydrothermal activity associated with active magma chambers.

### **1.1.3.2. Petrological studies of quenched materials tapped from magma chambers.**

Geochemical and mineralogical studies of those fine - grained rocks ( mainly lavas and ash - flow tuffs ) which represent quenched samples of high - melt - percentage magmas from magma chambers, and also of the cognate cumulate nodules which they sometimes contain ( Tait 1988 ), are mostly used to investigate geochemical processes rather than the physical processes which produce them. It is sometimes possible to identify the operation of physical processes, such as differential crystal settling, directly from chemical and mineralogical data ( Cox & Mitchell 1988, Sparks 1988 ) but in general studies of this type are most useful when a definite sequence of eruption of the quenched rocks can be correlated with spatial or temporal variations in the magma chamber whence they came. A detailed knowledge of the dynamics of magma withdrawal from magma chambers is needed to do this with confidence and such knowledge is at present largely lacking. Nevertheless, one of the most important results of this work has been the discovery of physical and compositional zonation in the magma chambers which fed ash - flow tuff eruptions ( Williams 1942 ). This zonation is believed on the basis of physical arguments to correspond to vertical stratification in the source magma chamber ( see 1.1.3.4. below ). A particular advantage of the quenched nature of these rocks has been that this permits the use of detailed mineralogical studies to estimate temperatures, volatile activities and crystal contents in the source magma chamber ( e.g. Druitt & Bacon 1988 ). When information is available on the precise ages of sequences of recent or sub - recent eruptive rocks, for example from historical records or precise isotopic dating, the study of these rocks can also provide information on rates and timescales of magma chamber processes. However, the essential limitations of all of these studies are the loss of information about the geometry of the source magma chambers and the distributions of magmas within them, and the almost complete lack of samples of high pre - eruption crystal contents ( above about 60% , Marsh 1981 ), which are inherent in the eruption process.

### **1.1.3.3. Field, petrographic, mineralogical and geochemical studies of large intrusions.**

Petrological and related studies of large intrusions, particularly of those of mafic and ultramafic compositions, have long been a favoured method of research into magma chambers and magma chamber processes. The origin of the considerable research effort put into them was the recognition by N.L.Bowen ( 1917, 1927 ) and others that many plutonic intrusions, particularly those of mafic, ultramafic and anorthositic compositions, are to a large extent made up of crystals separated from

magmas during crystal fractionation and do not correspond themselves to any liquid composition. The value of the corollary of this, that such intrusions were once sites of crystal fractionation and associated magma chamber processes, was first fully demonstrated by the work of L.R. Wager and W.A. Deer on the Skaergaard intrusion ( Wager & Deer 1939 ). Wager realised that the layering and other quasi - clastic sedimentary structures in Skaergaard represented evidence for the operation of crystal settling, density currents deposition and other sedimentary processes in the magma chamber which the intrusion had once been and that the cryptic variation in mineral compositions up the layered series showed that these processes were responsible for progressive crystal fractionation in the magma chamber.

The essential advantage enjoyed by this and subsequent studies is that one is actually looking at the sites of operation of the magma chamber processes being investigated. It is therefore possible to determine the size, geometry, heat loss regime and structural evolution of the magma bodies in which they took place, as well as being able to examine a vast variety of textures and structures produced by them. There are, however, two major problems associated with this approach. Firstly, the rates of magmatic processes such as crystallisation and formation of cumulates cannot be measured directly and have to be estimated indirectly via equations relating them to processes - most notably heat loss ( Irvine 1970 ) - whose rates can be calculated theoretically on the basis of other observations. Secondly, the intrusions do not directly correspond to magma chambers, but are instead the products of a prolonged period of crystallisation from them. This has a number of consequences of which the most obvious is that the intrusions are in no way a 'frozen in' image of the state of the magma chamber at a single point in time. Another consequence which has been appreciated for a long time is that the estimation of the compositions of the magmas present at any one stage, which is important because of the compositional control of magma rheology and density, and hence of patterns of convection ( McBirney 1985 ), is not a simple matter. Various procedures have been used which are based on calculating the effect of extracting the observed cumulate assemblages from an initial magma corresponding to the composition of chilled marginal facies, associated minor intrusions or pillow complexes within the intrusions ( Wiebe 1988 ): see Wager & Brown ( 1968 ), Morse ( 1981 ) for examples of these calculations. The problem with these methods is that they are critically dependent on the choice of initial composition, and if the rocks used are not representative, because of local contamination, alteration, residual melt expulsion, or subsequent replenishment of the magma chamber with liquids of differing compositions, the calculations and conclusions based

on them about the likely behaviour of the magma chamber will be invalid ( Hunter & Sparks 1987 ). Qualitative inferences, such as the detection of magma stratification through discordances between modal layering and cryptic variation ( Wilson & Larson 1985 ) or identification of magma inputs to the chamber during crystallisation, tend to be more secure. The input of magmas to a chamber can be identified by reversals in cryptic variation ( Brown 1956, Irvine 1980 ) or by changes in isotopic compositions ( e.g. De Paolo 1985 ) in the case of fresh inputs of primitive magma, or isotopically, in the case of crustal contamination ( Palacz & Tait 1985 ).

A more subtle problem caused by the prolonged crystallisation histories characteristic of plutonic rocks is that they tend to have been affected by a great variety of post - accumulation ( or post - cumulus ) and subsolidus recrystallisation processes such as infiltration metasomatism ( Irvine 1980 ) caused by compaction and interstitial melt expulsion ( McKenzie 1984 ), interstitial melt convection ( Sparks et al. 1985 ), textural equilibration ( Vernon 1970; Hunter 1987 ) and zone refining ( D.Walker et al. 1988 ). Interpreting the compositions, mineralogies and textures, and even mesoscopic structures such as layering in terms of the same phenomena as affect the compositions of magmas erupted from magma chambers is therefore not an easy procedure.

The rocks in large mafic intrusions which are least affected in this way, and are therefore most easily interpreted in terms of magma chamber processes, are the fine - grained rocks found in the marginal zones of many subvolcanic and other upper crustal intrusions. In addition, these rocks will have formed in the thermal and mechanical boundary layers which, according to recent theoretical studies ( see below ), control magma chamber behaviour and are also the likeliest sites of wall rock assimilation. Despite this, geological studies of these marginal zones are rare: notable ones include two studies of the margins of the Rhum complex, Dunham ( 1964 ) and Greenwood ( 1987 ).

Two other geological disciplines which have been applied to research into magma chambers, but rarely in conjunction with igneous petrological studies, are structural geology and metamorphic petrology. In contrast to the situation in granite petrogenesis, where the pivotal role of the 'space problem' in the Granite Controversy of the first half of this century has led to numerous structural studies of granites, field based structural studies of mafic and ultramafic plutonic intrusions are virtually non - existent, despite the information which they can provide on the replenishment of magma chambers ( Wiebe 1988 ) as well as their initial emplacement. Studies of the contact

metamorphism and hydrothermal alteration associated with intrusions can potentially be used to determine mechanisms and, through careful thermal modelling ( Parmentier & Schedl 1981 ), rates of heat loss from magma chambers. As noted above, this is perhaps the most promising approach to determining the rates of magmatic processes in ancient intrusions.

#### 1.1.3.4. The application of physics to magma chamber studies.

Theoretical studies, including analogue experiments in the laboratory using aqueous solutions, gels and oils as analogues for silicate melts, and numerical experiments using very powerful computers, have provided the stimulus for much of the recent work on magma chambers through the application of fluid dynamics and other branches of physics to several problems in the field ( see Turner & Campbell 1986, Huppert & Sparks 1984 for reviews of this field ). It is important to realise that these studies are **not** a source of empirical data on the behaviour of real magma chambers, but are instead a source of ideas and models with which to interpret empirical observations and a means of testing theories based on these observations for physical plausibility.

Probably the most important application of fluid dynamics to high melt - percentage magma chambers deals with patterns of convection in them, and in particular the development of compositional and thermal stratification. Numerical and, especially, analogue experiments have shown how stratified convection systems can be produced, by processes such as floor crystallisation producing denser residual melts, sidewall crystallisation, wall - rock melting, or replenishment with denser magmas; or destroyed, by replenishment with buoyant magma, cooling at chamber roofs, or floor crystallisation producing buoyant melts ( Huppert & Sparks 1980, McBirney et al. 1985, Turner & Campbell 1986, Huppert et al. 1986 ). Whether these experiments are actually relevant to real magma chambers is very dependent on the actual effect of processes such as fractional crystallisation on melt densities ( McBirney 1985 ). Another problem in theoretical studies of convection in magma chambers has been the selection of appropriate boundary conditions from which to estimate the vigour of convection, quantified by the Rayleigh number,  $Ra$ , of the flow; in other words, deciding what is actually driving the convection. A common procedure has been to treat magma chambers as fluid layers cooled from above and heated from below ( e.g. Martin et al. 1987 ), whereas in fact it is probably more realistic to estimate  $Ra$  from the rate of heat loss through the wall - rocks and consequent cooling in a thin boundary layer ( Brandeis & Jaupart 1986, Carrigan 1987, Marsh 1988 ): the two estimates

give values of  $Ra$  differing by many orders of magnitude! Similar problems arise in the analysis of compositional convection driven by density changes produced by fractional crystallisation instead of isochemical contraction on cooling.

Another problem which has been much influenced by theoretical fluid dynamic studies is that of the physical plausibility of the crystal settling mechanism proposed by Wager & Deer 1939, which has been challenged principally on the grounds that vigorous convection would keep the crystals in suspension ( McBirney & Noyes 1979 ). Numerical studies have shown that crystal settling can in fact occur in vigorously convecting magmas ( Weinstein et al. 1988 ), but other mathematical and analogue studies suggest that two other processes are much more efficient mechanisms for producing floor accumulation of cumulate rocks. The first of these involves density currents of crystal - laden magma, produced by sidewall cooling, sweeping down magma chamber walls in a manner analogous to marine density currents ( Wager 1963, 1968; Irvine 1987 ). The second involves similar dense 2 - phase plumes within the body of the chamber, produced by periodic instability and consequent detachment of the cooled and partially crystallised boundary layer at the roof of the magma chamber. This was first proposed by Grout ( 1918 ) but has recently been studied in quantitative mathematical models and shown to be a very efficient process indeed ( Brandeis & Jaupart 1986, Morse 1986, Marsh 1988 ).

Finally, theoretical studies ( notably McKenzie 1984 ) of gravity - driven compaction of, and melt expulsion from, crystal mushes ( first proposed by Bowen 1928 ) have profoundly influenced ideas on mechanisms of postcumulus crystallisation and adcumulate formation ( Hunter 1987 ). The theoretical studies predict that these processes may occur either in cumulate piles on the floors of high - melt - percentage magma chambers or in independent low - melt - percentage magma chambers ( Bedard et al. 1988 ), and also show that convection due to compositionally - controlled density variations may also occur in these low - melt - percentage rocks ( Sparks et al. 1985 ).

All of these theoretical studies have important limitations. The most fundamental of these is the uncertainty about whether or not the assumptions made and/or the chosen starting conditions are realistic. In the mathematical studies simplifications have to be made ( for example, about the variation in magma rheology with temperature ) in order to make the calculations tractable. Similarly, the fluids used in the analogue experiments have different fluid dynamic properties to

magmas, making it difficult to scale the experiments properly, and certainly cannot be made to simulate the crystallisation behaviour or temperature dependent rheology of silicate magmas. In any study of this sort there is always the problem that some natural phenomenon which has not been included in the mathematical or analogue models will change the results: it is therefore most important to look for evidence of such phenomena in natural rocks and intrusions.

In conclusion, a great variety of techniques exists for studying magma chambers from a distance ( either in time or space, or both ) as does a large body of theory for describing their physical behaviour. The principal problem facing the geologist studying actual rocks and/or magmas who wishes to utilise the theories to interpret his observations is that of gathering the data - for example, on rates of heat loss - necessary to apply the models. This may require the extension of his work to fields not normally considered part of igneous petrology.

## 1.2 Previous research on the Hypersthene Gabbro of Ardnamurchan Point.

### 1.2.1 Early Research and the work of the Geological Survey

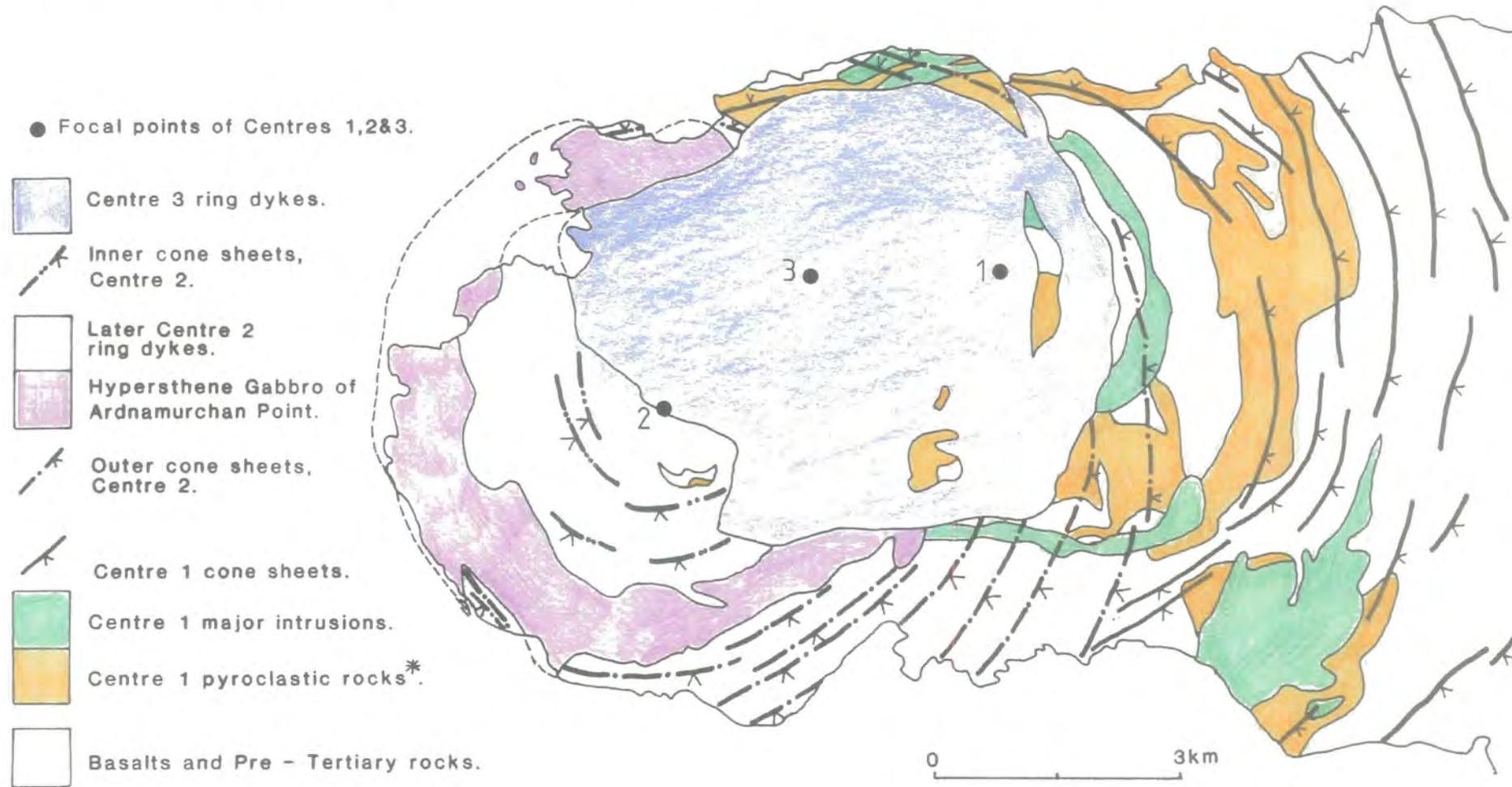
Although the Hypersthene Gabbro was not identified as a distinct pluton ( see note on intrusion nomenclature below, section 1.2.4 ) until western Ardnamurchan was mapped by Scottish Geological Survey workers in 1922 - 23 ( Richey et al. 1930 ), it first appears in the literature much earlier, in J.W.Judd's work on the central intrusive complexes and other igneous rocks of western Scotland ( Judd 1874 ). Judd noted that the gabbros making up the nose of the Ardnamurchan Peninsula were, in the western part of their outcrop, characterised by a high modal hypersthene content. He also noted that Jurassic sediments and other country rocks were strongly domed around the gabbros and were highly metamorphosed within a broad contact aureole. Judd's work is also of much wider significance in that he recognised for the first time that the intrusive and extrusive rocks of the Hebrides were broadly co - genetic and the plutonic rocks of the central complexes represented the deeply eroded cores of central volcanoes. He also realised that the central complexes could therefore be used to investigate the sub - surface components of volcanic activity ( Judd 1874, p.232 ).

The much more detailed work of Richey et al. ( 1930 ) established that the Hypersthene Gabbro of Ardnamurchan Point, as the newly defined pluton was named, formed an arcuate area of outcrop at the western end of the peninsula ( see Fig. 1.2 ). Richey and his co - workers considered that the intrusion had been much reduced in size by later intrusions and that its original geometry was that of a ring - dyke, some 8km by 6km in its outer dimensions and up to 1.5km thick ( both as measured in the horizontal plane ). They considered that its outer margin, where beneath the sea, was not far offshore because of the presence of a well - defined marginal facies in coastal outcrops. Their evidence for the position of the inner margin is described below. The present outcrop area is divided into two parts by Sanna Bay, but that the two parts belong to the same pluton has never been in doubt, because of the consistent age relationships shown by them, the presence of a quartz dolerite marginal facies and the unusual orthopyroxene - rich mineralogy of the rest of the pluton in both areas. This latter feature was recognised by Richey et al. ( 1930 ) as being unique within Ardnamurchan and highly unusual in the British Tertiary Igneous Province as a whole.

As the authors of the Ardnamurchan memoir themselves recognised ( Richey et al. 1930, p.55, 58 – 59 ), their interpretations of the Hypersthene Gabbro and of other arcuate and annular intrusions in Ardnamurchan were greatly influenced by previous Geological Survey work on central complexes, in Glencoe ( Clough et al. 1909 ) and in Mull ( Bailey et al. 1924 ) in particular. The Survey's previous work on the Tertiary central complexes had shown that they were, in general, divided into successive centres produced by slight shifts in the centre of activity with time. Each centre was defined both by its age relationships and by a crude concentric arrangement of the arcuate, annular and circular intrusions within it, as well as a radial arrangement of dykes, around a common centre or focal point. Two main types of concentric intrusion, hypabyssal cone sheets and larger, steep - sided annular ring dykes, were recognised. The latter are particularly associated with central subsidence, as in Glencoe and the Loch Ba area of Mull. Where multiple ring dykes had been found, they were separated by screens of older rock, which confirmed their geometry but prevented direct determination of their age relationships. A number of variations on the ring dyke theme were described by the Survey workers and others in the 1920's, notably flat - topped and bell - jar intrusions produced when the outward - dipping ring fracture defining the intrusion closes off at depth rather than intersecting the surface ( Richey 1932 ). A structural model for the formation of cone sheets and ring dykes was developed by E.M.Anderson, in which cone sheets were produced by tensile fracturing **above** an overpressured magma chamber and emplacement of magma along the fractures, whilst ring dykes were considered to be passive intrusions formed by foundering of parts of the magma chamber roof when the magma pressure in the chamber fell below that exerted by the weight of country rocks ( E.M.Anderson in Bailey et al. 1924, Anderson 1936 ). By the time the survey of Ardnamurchan began, the concept of subvolcanic central intrusive complexes as being made up of one or more centres defined by confocal suites of cone sheets, ring dykes and radial dykes had therefore attained the status of a paradigm, to which the Cuillins gabbro, regarded as a set of superimposed laccoliths ( Harker 1904 ), and the forcefully emplaced North Arran Granite, were regarded as unusual exceptions ( Richey et al. 1930, p.55 ).

The paradigm reached its most complete development with the work of Richey and his co - workers in Ardnamurchan: Centre 3 of the complex ( see Fig 1.3 ) in particular has long been considered to be the most perfect example of a centred suite of ring dykes in the British Tertiary Province. Richey et al. 1930 defined three igneous centres within the Ardnamurchan complex on the basis of both age relationships and intrusion distribution ( see Fig 1.3 ), and the overall chronology

**Fig 1.3. Subdivision of the Ardnamurchan Central Complex into 3 Intrusive Centres containing cone sheets and ring dykes, according to Richey et al. 1930.**



\*Agglomerates considered to be intrusive by Richey et al.

established by them has been accepted ever since. Within the three centres, Richey et al. were able to establish an unusually complete relative chronology because the intrusions are only very rarely separated by screens of older rocks, allowing determination of age relationships directly from the contact relationships. The penalty incurred by this, as was realised at the time ( Richey et al. 1930, p.203 - 204 ), is that the original geometry of individual intrusions is less easily ascertained, because the centres show an overall inward younging: with only a few exceptions, the inner edges of the intrusions are formed by younger intrusions. Under these circumstances Richey and his co - workers resorted to a number of indirect methods to infer that most of the intrusions were originally ring dykes, rather than stock - like intrusions.

In the case of the Hypersthene Gabbro, Richey et al. concluded that it was a ring dyke on the following grounds:

1). The outer contact dipped steeply outwards and was discordant to bedding in the country rocks, except in a few small areas, where it dipped outwards at lower angles and was roughly concordant to bedding, notably around Glebe Hill where hornfelsed basalts overlying gabbros were considered to be part of the roof of the pluton. The pluton was considered to be a 'blind' ring dyke, with the steep walls passing up into a domed roof, rather than intersecting the surface to form a caldera.

2). The incomplete annular or arcuate horizontal cross - section of the pluton was recognised to be consistent with it having once been a ring - dyke, once it was realised that the eastern part of the pluton could well have been removed by the emplacement of the intrusions of Centre 3 ( Fig 1.3 ).

3). Although the inner edge of the pluton is entirely formed by contacts with younger intrusions, Richey et al. ( 1930 ) describe fine - grained, fluxioned or flow - laminated rocks close to the inner contact which they considered to have formed by flow alignment close to the steep inner margin of the pluton.

A corollary of the interpretation of the Hypersthene Gabbro as a passively emplaced ring dyke is that its emplacement could not have produced the radial outward dips in the country rocks, which define a dome centred on Centre 2. This was recognised by Richey et al., and they argue that it is consistent with the age relationships which they deduced, as follows:

1). The doming of the country rocks was believed to predate the emplacement of the outer cone sheets of Centre 2 ( Fig. 1.3 ), because the latter do not show a decrease in their inward dip within the area of doming, relative to cone sheet dips outside it, as they would be expected to do if they predated the doming.

2). The outer cone sheets of Centre 2 are almost all apparently truncated by the Hypersthene Gabbro and are affected by metamorphism which Richey et al. attributed to the Hypersthene Gabbro ( see below ). These cone sheets were therefore considered to predate the emplacement of the Hypersthene Gabbro.

3). Cross - cutting relationships within Centre 2 indicated that the Hypersthene Gabbro was older than all the other ring dykes in the centre, except possibly the Old Gabbro of Aodann, with which it has indeterminate age relationships but from which it is distinguishable by its orthopyroxene - rich mineral assemblages.

Richey et al. therefore concluded that the doming must predate any of the large intrusions now exposed in Centre 2, and was probably produced by the initial emplacement at depth of the intrusion whose subsequent expansion resulted in the emplacement of the outer cone sheets.

Although they found no major contacts within the Hypersthene Gabbro and considered that it was best regarded as a single intrusion, the Survey workers distinguished two main suites of rocks within the Hypersthene Gabbro: a marginal facies dominated by quartz dolerite and a coarser suite, dominated by olivine gabbro-norites, in the interior of the intrusion. It was recognised that the outer fine - grained margin of the intrusion is unusually broad: this was interpreted as evidence of unusually rapid cooling of these rocks, although the cause of this was not identified. Richey et al. suggested that the quartz dolerites represented a slightly earlier and more evolved batch of magma than that of the main part of the intrusion. However, they considered that the transition between the two groups of rocks was gradational and therefore did not map the quartz dolerites as a separate intrusion. The presence of numerous acidic veins cutting the quartz dolerites was noted; these were considered to represent residual melts expelled from the interior of the intrusion. Similarly, heterogenous rocks with mafic primocrysts and a granophyric mesostasis were interpreted to be the product of 'soaking' of basic rocks in a migrating interstitial acid melt. Wall rock melting

and assimilation were not considered to have played a part in the formation of these rocks.

Richey and co-workers found that the interior of the intrusion was dominated by relatively coarse grained ophitic olivine gabbro-norites with roughly equal proportions of ortho- and clinopyroxene, but showing modal variation between near-troctolitic rocks and olivine-poor gabbro-norite. This variation was found to be non-systematic; it was recognised that the interior of the intrusion was unzoned. In addition to the modal variation in the coarse grained rocks, a number of finer grained sheets and xenolithic strips were found during the course of the survey mapping. These were all characterised by the development, to varying degrees, of granoblastic hornfels textures but varied widely in type and inferred origin. Most were considered to have been derived from mafic igneous protoliths, including hornfelsed microgabbro sheets, annealed cataclastic shear zones, dolerite and basalt xenoliths, and banded rocks which were believed to be the products of flow deformation, but some anomalous xenoliths were found. The best-known of the latter are the hercynite-plagioclase-corundum-cordierite xenoliths from the north side of Glebe Hill, which were interpreted as meta-laterites (H.H. Thomas in Richey et al. 1930). Thomas also compared the basic hornfels sheets with similar rocks in the Cuillins gabbros of Skye (Harker 1904) and in the Ben Buie gabbro, Mull (Thomas in Bailey et al., 1924). Thomas considered that the granoblastic microgabbro sheets in both Mull and Ardnamurchan represented spalled fragments of early facies of the gabbros which had been hornfelsed by, and equilibrated with, later magmas in the same intrusion, with the implication that they should predate their hosts.

The Survey workers also re-investigated the metamorphic aureole around the intrusion which had first been identified by Judd (1874). To their surprise, in view of the evidence for rapid consolidation and cooling of the margin of the intrusion, the aureole was found to be unusually broad and to contain high-grade mineral assemblages (Richey et al. 1930, p.217,235). The following assemblages were noted (see map 1 for localities):

- 1) Pelitic hornfelses in the Dubh Chreag area, within 200m of the contact: Biotite-hypersthene-cordierite-magnetite bearing rocks with abundant short plagioclase laths set in an intergrown quartz-alkali feldspar mesostasis. H.H. Thomas compared these to buchites and considered that they had almost melted.

2) Calcareous shales, also from Dubh Chreag: augite - actinolitic hornblende - calcic plagioclase - sphene. Tremolite and garnet were recorded from purer calcsilicates at Sron Bheag.

3) Basalts, from between Druim na Gearr Leacainn and Glebe Hill: these were found to show progressively coarser and more complete recrystallisation towards the contact, with the appearance of biotite being followed by development of granoblastic textures and finally the appearance of orthopyroxene.

4) The Cone Sheets: these were found to show some recrystallisation, but, unlike the basalts, largely retained primary textures including chilled margins. The characteristic metamorphic reactions were identified as the replacement of secondary chlorite by biotite, recrystallisation of augite, replacement of primary feldspar by albite, and recrystallisation of opaques.

Pre - metamorphic cataclasis, followed by syn - metamorphic annealing, was identified in both the basalts and the cone sheets. H.H.Thomas suggested that this deformation might have been produced by the emplacement of the Hypersthene Gabbro, but the age relationships of this deformation were not systematically investigated by the Survey workers. Thomas did not attempt, either, to determine whether the assemblages noted above were stable equilibrium assemblages, or to consider the implications of the different assemblages developed in the basalts and in the cone sheets.

#### **1.2.2. The Complete Re - Investigation of the Hypersthene Gabbro by M.K.Wells**

The interpretation of the plutonic rocks of Ardnamurchan as classic or paradigmatic examples of ring dyke complexes has been emphasised in the above account of the work of the Geological Survey on the Hypersthene Gabbro because most of the work done on it since the 1930's has been carried out with the purpose of testing or revising the ring dyke model for its emplacement. This work began with that of M.K.Wells ( Wells 1951, 1954 ), which started as an investigation of the geometry and emplacement of the intrusion but broadened into a more general study covering its petrology and geochemistry as well ( Wells 1954 ).

Wells rejected the ring dyke model for the emplacement of the Hypersthene Gabbro on the basis of three main lines of evidence ( Wells 1954 ). Firstly, he identified compositional similarities between

the marginal quartz dolerites and the outer cone sheets of Centre 2 ( Fig 1.3 ), and proposed that they originated from the same batch of magma. He interpreted Anderson's model ( Anderson 1936 ) for cone sheet formation as implying that the doming and the cone sheets formed in the same stress field, above a spheroidal body of overpressured magma, and therefore proposed that the doming was produced during the upwelling of the magmas which subsequently formed the Hypersthene Gabbro. However, Wells accepted Richey's contention that the doming predated the emplacement of the cone sheets and, therefore, the final emplacement of the Hypersthene Gabbro.

Secondly, Wells remapped the outer contact of the pluton and concluded, largely on the basis of structural contouring of the contact in areas of poor exposure but considerable topographic relief, that it was broadly conformable to the outward - dipping country rocks. In doing so he was obliged to make the assumption that, except where it was cut by large faults, the contact formed a smooth conical surface. Some outcrops where the contact was steeply dipping were noted by Wells but these were considered to be anomalous. Wells also identified areas at or close to a posulated flat - lying pluton roof, respectively around Glebe Hill and on Druim Reidh - Dhalach. In the latter case the proximity of the roof was inferred from the presence of a large area of hornfelsed microgabbro which Wells considered to be a roof facies. In addition to, and at variance with these observations, Wells also identified a large number of basic hornfels strips and sheets within the Hypersthene Gabbro which he considered to be intensely metasomatised xenoliths of metasedimentary, metabasaltic, or metadoleritic rock. He accounted for the uniform orientation ( dipping toward Centre 2 at 30 to 60 degrees in sub - parallel swarms ) of these bodies by proposing that they had formed more or less in situ and that the host gabbro norites had been emplaced by a process of wedging - open of inward - dipping fractures in the country rocks. This implied that the contact dipped **inwards**.

Thirdly, Wells re - interpreted the layering and banding within the Hypersthene Gabbro, which Richey et al. ( 1930 ) had considered to be flow banding parallel to the inferred inner wall of the intrusion, in terms of the floor accumulation model originally proposed by Grout ( 1918 ) and greatly developed by Wager & Deer ( 1939 ). Wells recognised that, although the layering was patchily developed, and there was no systematic cryptic variation, it did have a systematic pattern of orientation. He found that the layering was concentric to Centre 2 and showed an inward increase in dip from 10 to over 50 degrees toward Centre 2. This was considered to be a primary feature.

Using this evidence, Wells proposed that the pluton had had a circular horizontal cross - section prior to the removal of its core and eastern margin by later intrusions. He also suggested that it had a domed, broadly conformable upper surface and an inverted - conical lower surface with marked upward flare, similar to that shown in Fig 1.4.

Wells' more general petrological studies of the Hypersthene Gabbro revealed a variety of rock types in its interior in addition to the dominant olivine gabbro, notably coarse troctolites and a variety of coarse and pegmatitic augite - rich gabbros. He was in general unable to map discrete intrusions of these rocks, finding instead that those intrusive contacts that were present were discontinuous and often passed laterally into gradational transitions. The principal exception to this rule was provided by a number of beerbachitic ( granoblastic dolerite; Phillips 1969 ) dykes and veins; Wells considered these to be rare and that the vast majority of the fine grained granoblastic rocks within the pluton were xenolithic in origin.

During this work on the Hypersthene Gabbro, Wells was greatly influenced by H.H.Read's ideas on metasomatic granitisation ( Wells 1951,1954 ). This is particularly apparent in his work on the marginal quartz dolerite and also the adjacent country rocks in the area southwest of Rubha Carrach ( Grid square 4670; see also Map 2 and Fig 3.1 ). A radical metasomatic model was proposed for the rocks near Rubha Carrach, involving differential diffusion of ions into country rocks which were postulated to have originally been pure quartzites. At its most intense, near the contact, this process was believed to have produced 'pseudobreccias' with relict blocks of quartzite in a microgranitic matrix. The collapse of the granitisation theory ( Bowen 1948 ) invalidates Wells' interpretation of these rocks, whose origin is discussed further in Chapter 3 and section 4.2.4. Like Richey et al. ( 1930 ) Wells considered that the marginal quartz dolerite has a gradational contact with the rest of the pluton. He also believed that it had been extensively modified by the outward migration of volatile - rich melts from the crystallising interior of the intrusion, producing its mottled appearance and the extensive network of granitic veins which cuts it ( Wells 1954 ). Wells also noted extensive hydrothermal alteration of the marginal rocks, which he also attributed to migration of hydrous volatiles from the interior of the pluton. He did not consider that local crustal contamination was involved in the formation of the quartz dolerites, because of the lack of correlation between the composition of the quartz dolerite and the varied country rock lithologies. However, Wells found, within the interior of the intrusion, a number of banded xenoliths which he

considered to be metasomatised laminated sediments ( Wells 1951 ): Brown ( 1954 ) pointed out that these could be xenolithic blocks of layered cumulate rocks and drew analogies with similar rocks in Rhum.

### **1.2.3. Further re - interpretations of the structure of the Hypersthene Gabbro and the application of some new techniques.**

In contrast to Wells ( 1954 ) and in agreement with the earlier conclusions of Judd and of Richey and his co - workers, Skelhorn & Elwell ( 1971 ) considered the Hypersthene Gabbro to be a steep - sided discordant intrusion. They recognised that the crystal accumulation model for the formation of igneous layering does not allow one to infer the geometry of an intrusion from the orientation of the layering within it ( contrast Wells 1954 ) and suggested instead that the very steep inward dip of the layering close to the inner margin of the intrusion owed its orientation to post - depositional drag folding around the subsiding core of the intrusion, now buried beneath the present level of exposure. They suggested that this central subsidence was associated with the emplacement of the Sgurr nam Meann hybrid dolerite ring dyke ( intrusion 1 of this rock type in Fig. 1.2 ). Skelhorn and Elwell therefore concluded that the intrusion had originally been a steep - sided plug - like pluton. At about the same time, Wells & McRae ( 1969 ) examined a similar model and showed from the palaeomagnetism of the pluton that, if folding of the layering had occurred, it must have done so prior to the magnetisation of the rocks as they cooled through their Curie temperature of about 570°C.

A more radical re - interpretation of the Hypersthene Gabbro proposed in recent years stems from Wells' observations of sub - parallel granoblastic basic sheets, which he had interpreted as screen - like in situ xenoliths, and of internal contacts within the pluton ( Wells 1954; section 1.2.2 ). These observations suggested that the pluton might have been built up over an extended period by a large number of small magma injections into a hot, partially molten pluton built up from previous small magma batches, an idea that was subsequently developed into an interpretation of the Hypersthene Gabbro as a confluent cone sheet complex ( G.P.L. Walker 1975, Wells 1978 ). Confluent cone sheet complexes were first proposed by Richey ( 1932 ), who used the Cuillins complex of Skye as the type example. They are regarded as being made up of contiguous or confluent sequences of confocal cone - sheet - like intrusions arranged around an axial magma conduit and showing a marked inward

increase in dip ( in which they differ from classical cone sheets ). Individual sheets are considered to be emplaced with such frequency that the complex as a whole remains hot throughout its period of formation. This rapid sequence of emplacement is required to explain the lack of chilled margins and the coarse grain size characteristic of the Cuillins complex, the Hypersthene Gabbro and other proposed confluent cone sheet complexes. Only intensely metamorphosed slivers of country rock and early chilled cone sheets are postulated to remain between the later confluent sheets. According to Wells ( 1978 ) these account for most of the granoblastic sheets described by Wells ( 1954 ) but the later paper also notes the presence of unchilled but fine - grained moderately granular basic sheets which were inferred to have been emplaced whilst the host gabbro was still hot.

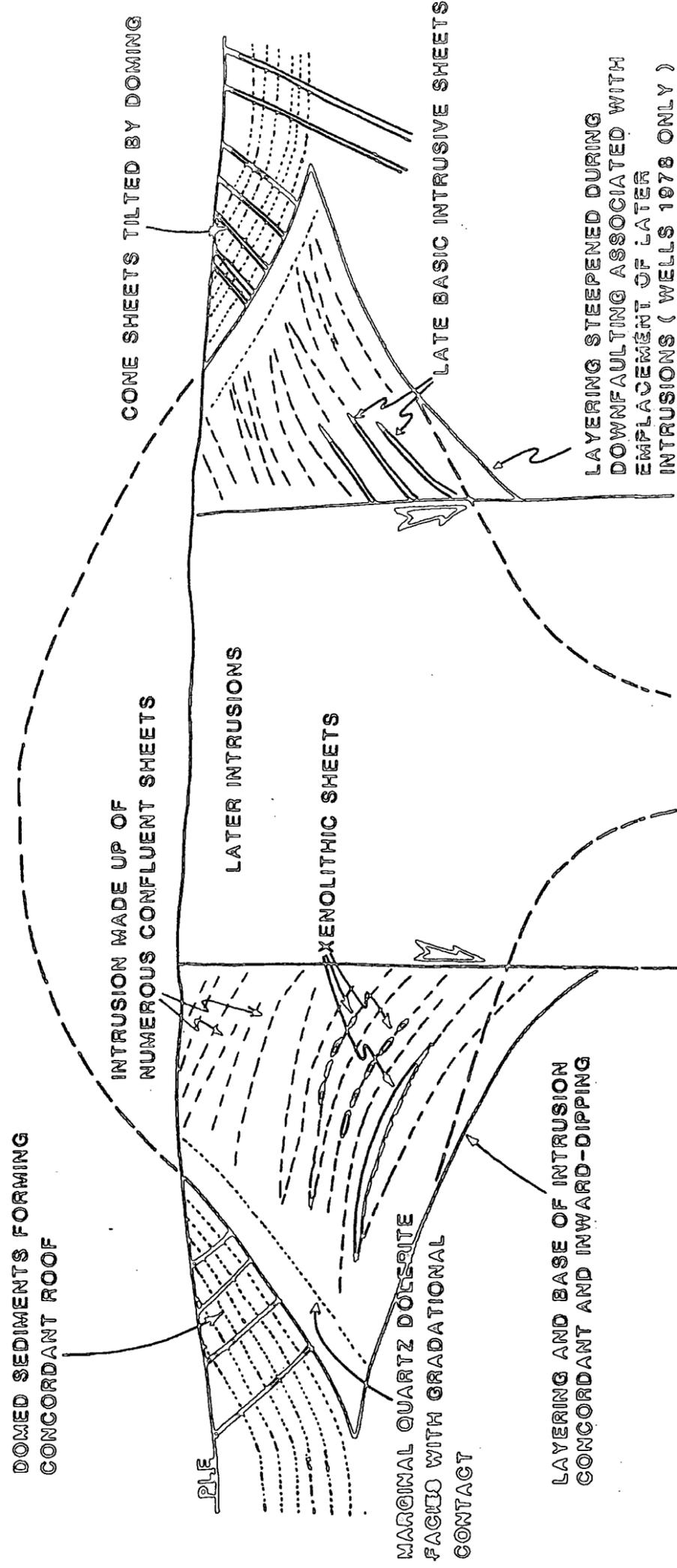
The two models of this type proposed for the emplacement of the Hypersthene Gabbro ( G.P.L. Walker 1975, Wells 1978 ) differ in some respects. Both predict that the upper and lower outer contacts of the Hypersthene Gabbro, and therefore of the plutonic component of the Ardnamurchan complex as a whole, should be, respectively, shallowly inward- and outward- dipping. Wells' model also implies that it should be concordant to the overlying roof rocks. With regard to emplacement mechanisms, Wells ( 1978 ) considers that space for the intrusion was created by doming of the roof rocks, and asserts ( in contradiction to Wells 1954 and Richey et al. 1930 ) that, as required by this model, the earlier cone sheets within the region of doming were tilted outwards. G.P.L. Walker ( 1975 ), in contrast, suggests that the doming was caused by the rise of an earlier acid diapir and that the outer Centre 2 cone sheets and the Hypersthene Gabbro itself formed as a result of the deflection of the hydrostatic stress field around the low - density acid rocks. The implications of the two models for the geometry of the intrusion are summarised in Fig. 1.4.

In contrast to the attention paid to its structure in recent years, essentially no petrological work has been done on the Hypersthene Gabbro in recent years, with the important exception of the work of Forester & Taylor ( 1971 ) on  $^{18}O$  depletions in the subvolcanic complexes of Skye, Mull and Ardnamurchan. This indicated that, like all the other intrusions in the area, the Hypersthene Gabbro had been affected by circulating low -  $^{18}O$  meteoric groundwater as it cooled.

#### **1.2.4 Advantages of the Hypersthene Gabbro as a 'natural laboratory' for the study of subvolcanic magma chambers**

Fig. 1.4. Two interpretations of the Hypersthene Gabbro as a confluent cone sheet intrusion ( not to scale ).

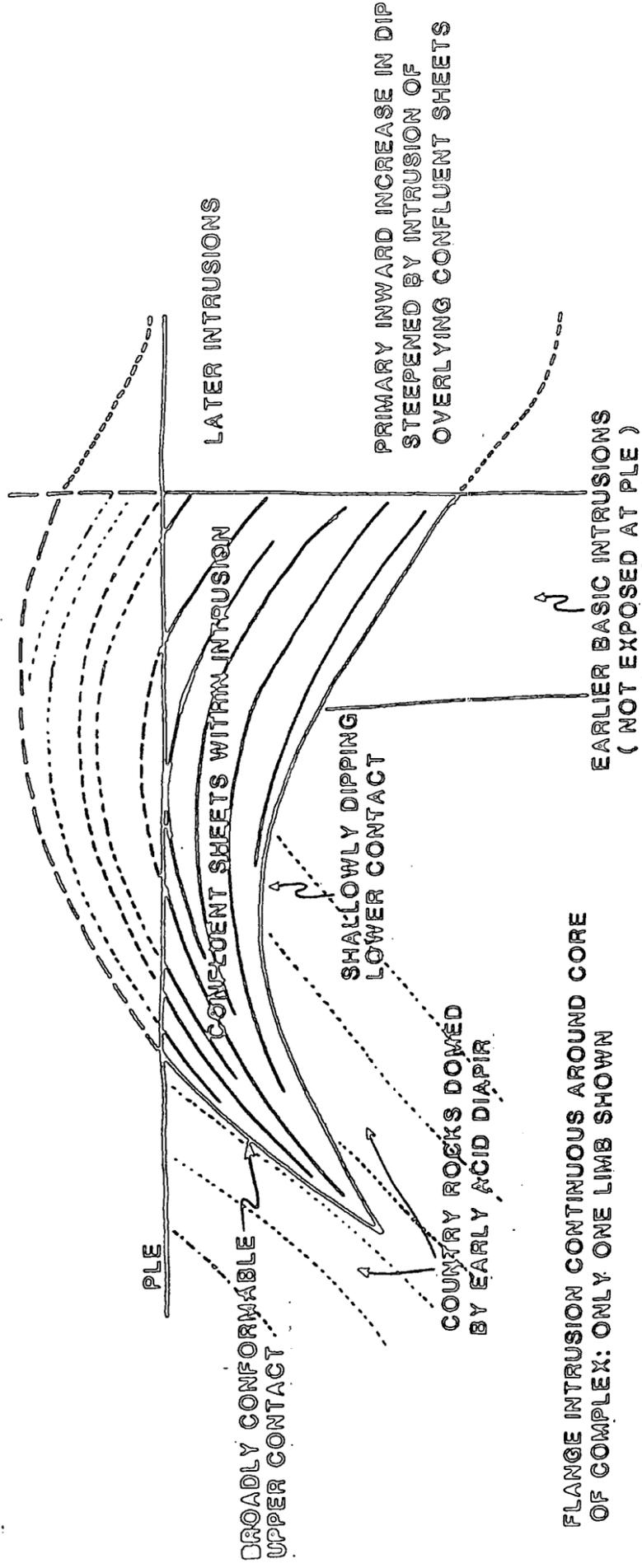
1. Wells ( 1954, 1978 ): cartoon depicting main features of Wells' model in cross-section.



PLE = PRESENT LEVEL OF EXPOSURE

ORIGINAL AXIAL CONDUIT

2. Walker ( 1975 ): The Hypersthene Gabbro as a flange intrusion formed by subsidence around dense basic intrusions beneath an acid diapir.



The work in this thesis falls firmly within the third type of study of magma chamber processes defined above ( section 1.1.3 ). As far as this type of study is concerned the Hypersthene Gabbro satisfies the necessary requirements of reasonably abundant exposure and high - quality outcrop ( having been heavily glaciated ), although the vertical extent of outcrop in particular is not as great as could be desired. It also has several more specific advantages which are apparent from the previous work described above. These can be summarised as follows:

1) The broad marginal zone of unusually fine grained rocks, noted first by Richey et al. ( 1930 ), suggested that evidence of boundary layer phenomena ( sections 1.1.3 & 1.1.4 ) might be preserved there.

2) The heterogenous nature of the marginal rocks, also first noted by Richey et al., suggested, on more recent interpretations of similar rocks elsewhere ( e.g. Dunham 1964 ) that processes such as magma mixing might have taken place there, and could be studied at outcrop and by means of geochemical analysis.

3) The high grade contact metamorphism noted by Richey et al., and possible local partial melting ( Thomas in Richey et al. 1930 ) suggested that the pluton would be a particularly good site for the study of wall - rock melting and assimilation, because of the unusual combination of a high - grade contact aureole and fine grained, rapidly solidified marginal rocks.

Some of the above observations by previous workers were confirmed during previous work by the author in the course of an undergraduate mapping project in northwestern Ardnamurchan ( Day 1985 ), in particular the presence of partial melting in the country rocks.

It was therefore apparent that the pluton would be particularly suitable for the study of boundary layer processes and their geochemical consequences, such as crustal contamination and magma - mixing. The pluton also proved to be amenable to studies of heat loss and its effects on magma chamber behaviour, because of the evidence provided by the well - developed contact aureole first noted by J.W.Judd.

#### **A note on nomenclature.**

Owing to the controversy over whether the Hypersthene Gabbro was emplaced in a single intrusive event or in a large number of closely - spaced events ( section 1.2.3 ), referring to it in its entirety as an intrusion is apt to prove confusing. In this work, therefore, the Hypersthene Gabbro as a whole will be referred to by the non - committal term **pluton**, whilst the use of **intrusion** will be restricted to bodies believed on secure grounds to have been emplaced in single events.

Another point of nomenclature which is dealt with at this point concerns the use of the terms phenocryst, porphyrocryst and xenocryst. The first and last have genetic connotations ( respectively, an early - formed crystal which grew in the magma ( or in one of the components of that magma, if it is of hybrid origin ), and a crystal derived from an unrelated rock ). Some of the problems to be discussed in this thesis revolve around the question of whether particular large crystals are phenocrysts or xenocrysts and the purely descriptive term porphyrocryst is used when describing the crystals, prior to identification of their origin as magmatic, xenocrystic or restitic.

The name **Hypersthene Gabbro of Ardnamurchan Point** is also problematic in that it is too specific a petrological description for modern usage, is incorrect in that most of the orthopyroxene in the intrusion is in fact bronzitic ( Chapter 6 ), and is misleading in that in modern usage the use of a mineral name as a prefix to a rock name implies that that mineral is an accessory phase only. However, for reasons of priority and continuity, the name will be retained in this work as the proper name of the **pluton** , whilst modern usage will be followed in assigning names to **lithologies** .

## 2. The Pre - Hypersthene Gabbro geology of Ardnamurchan: The Distribution and Composition of potential Contaminants, the Previous Structure of the area and Structural Markers.

The geology of those rocks in Ardnamurchan which predate the Hypersthene Gabbro is relevant to the evolution of the latter, and to the understanding of the magmatic and other processes which operated within and around it, for a number of reasons. The most important of these is that the older rocks provide potential sources for a range of crustal melts and contaminants of varied compositions. As will be seen, many of the pre - Hypersthene Gabbro rock types have well constrained spatial distributions as well as distinctive compositions. They can therefore be used not only to establish the fact of crustal anatexis or of contamination of magmas, but also constrain where this took place. A second relevant aspect of the country rock geology is the pre - Hypersthene Gabbro structure of the area, which needs to be understood because of its effect on the distribution of country rock lithologies and also because of the possibility that the emplacement of the intrusion may have been affected by pre - existing structural anisotropies. The most obvious of these latter are faults and shear zones, but bedding and schistosity, which will promote deformation of the rocks by flexural slip, may also have an effect upon emplacement mechanisms ( Pollard & Johnson 1973 ). Finally, correlations of distinctive lithologies within the country rock sequence across areas of poor or no exposure allow deduction of the large - scale structures in the country rocks and hence the pattern of deformation associated with the pluton. The use of the country rocks in this way requires that any pre - existing lateral variations, due to earlier deformation or sedimentary thickness or facies variation, be recognised and allowed for.

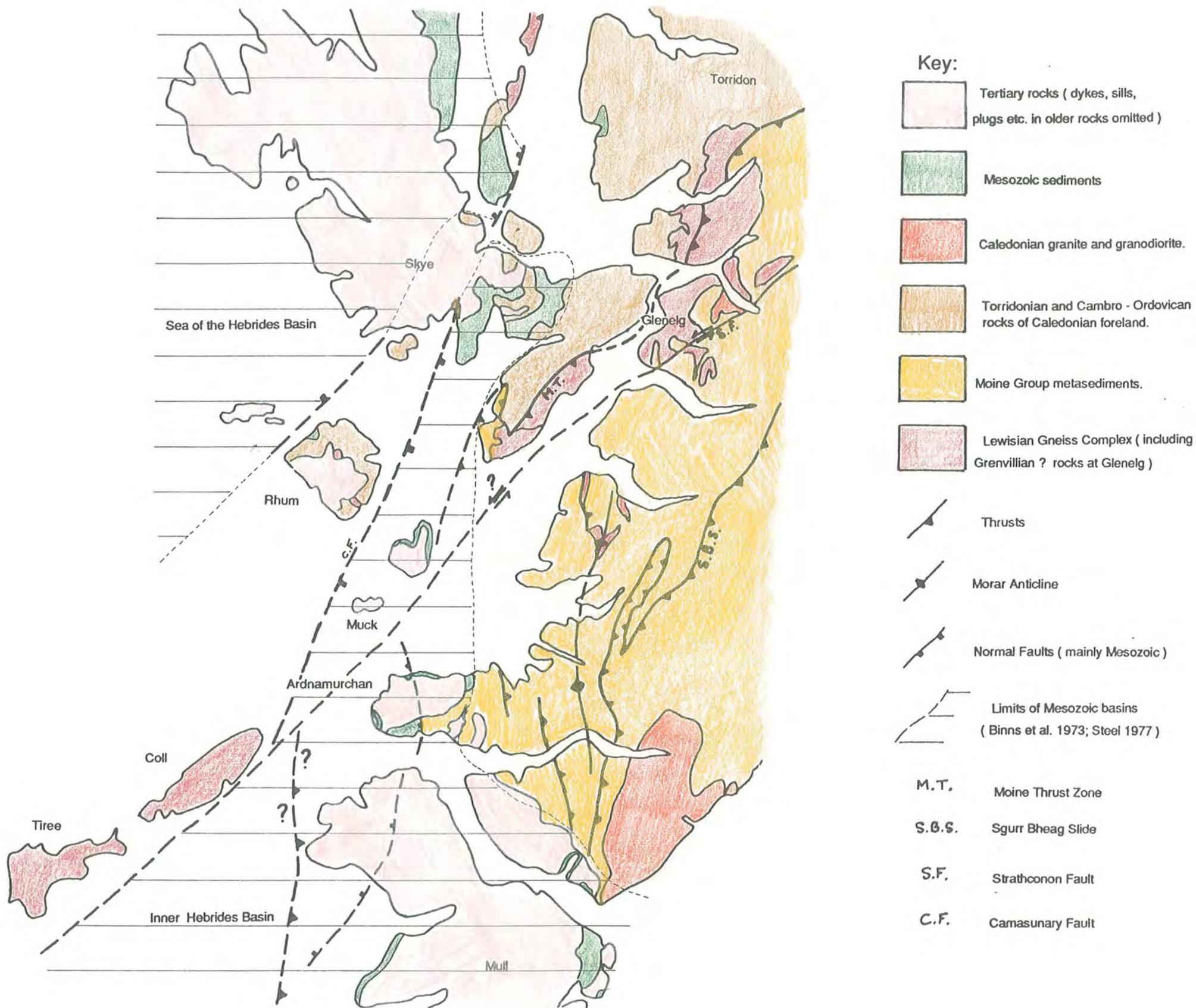
The groups of rocks covered in this chapter are, in order of decreasing age, Lewisian gneisses; Moine Group metasediments; Mesozoic sediments; and earlier Tertiary basalts, volcanoclastic rocks and intrusions. It should be noted that although the 'Outer Cone Sheets of Centre 2', which form about a quarter of the outer contact of the Hypersthene Gabbro by area, have in the past been considered to significantly predate it ( Richey et al. 1930; Wells 1954 ), evidence suggesting that the vast majority, if not all, of these sheets were intruded after the initial emplacement of the magma chamber corresponding to the Marginal Border Group of the pluton ( see sections 3.1 and 3.3 for a discussion of a revised subdivision of the pluton ) is presented in Chapter 3 ( sections 3.2.2 and 3.2.5 in particular ). The cone sheets will therefore be dealt with in Chapters 3, 4 and 7.

## 2.1. The Lewisian Complex.

Late Archaean and early Proterozoic granulite- and amphibolite - facies metamorphic rocks of the mainly meta - igneous Lewisian complex are generally considered to form the middle and lower crust of much of northern Scotland ( Smith & Bott 1975; Bamford et al. 1977 ), as well as outcropping at the surface to the north and west of the Moine thrust and as a number of inliers within the Caledonian fold belt ( Fig. 2.1 ). These latter are, however, relatively small in area and it seems likely that the top of the Lewisian complex beneath Ardnamurchan and much of the rest of the North - west Highlands of Scotland is, to a large extent, formed by the Moine thrust ( see section 2.2. ). Although Bamford et al. detected a relatively well - defined seismic boundary beneath the Central Highlands which has been interpreted as the transition from Lewisian to late Proterozoic ( Grenvillian? ) crust to the south and east, the dating of the Glenelg inlier as Grenvillian by Sanders et al. ( 1984 ) suggests that younger Proterozoic rocks may form part of the lower crust beneath the North - west Highlands. The Glenelg inlier is, however, composed of basic garnet - granulites and eclogites which have refractory bulk compositions. This suggests that they are unlikely to be important crustal contaminants of any but the most primitive, high temperature Tertiary magmas even if similar rocks are important constituents of the middle and lower crust in the area. This argument is consistent with the results of radiogenic isotope work on rocks of the British Tertiary province, particularly Pb isotope studies ( Dickin 1981; Dickin et al. 1981; Dickin et al. 1984 ) which have only ever detected Grenville - age contaminants in Arran, to the south of Bamford et al.'s seismic boundary. It should however be noted that the relevant isotopic data do not exist for any of the rocks in the Ardnamurchan complex and the possibility of contamination of Ardnamurchan rocks with Grenville - age crust cannot be excluded. For the purposes of the present work, however, the important distinction is between the mainly meta - igneous crust beneath the Moine thrust and the mainly meta - sedimentary rocks above it. Since the sub - Moine Thrust Zone rocks are not exposed in Ardnamurchan the detection and modelling of crustal contamination of Tertiary magmas by them has to rely on compositional data for equivalent rocks from adjacent areas.

The closest outcrops of basement rocks to Ardnamurchan are the Lewisian gneisses of Coll and Tiree, two islands west of Ardnamurchan. Although only 15km from Ardnamurchan at their closest point, these islands are on the far side of the Strathconon fault, a major Caledonian - age structure which may be a sinistral strike - slip fault with as much as several tens of kilometres of displacement across

## 2.1 Pre - Tertiary Geology of the central Inner Hebrides



it. However, provided that the displacement was less than fifty kilometres, the effect of movement on this fault would actually be to bring the islands into closer proximity to Ardnamurchan. They are thus the southernmost and nearest occurrences of Lewisian rocks for which major and trace element data are available ( Drury 1972, 1974, 1978; Tarney et al. 1979 ).

Coll and Tiree were first mapped by Richey et al. ( 1930 ), who found that they were mainly formed by tonalitic to granodioritic amphibolite - facies gneisses similar to the 'grey gneiss' of much of the Lewisian. About 10% of the outcrop is formed by metasedimentary gneisses, and even smaller areas of metabasic rocks, granite gneisses and post - metamorphic granites and granitic pegmatites were also found. The paragneisses were found to include quartzites, semi-pelites, calc-pelites, calc-quartzites and marbles, but true pelites are absent. Small areas of granulite - facies gneisses were found in both Coll and Tiree by Richey et al., and retrogression to greenschist facies, chlorite - prehnite - epidote bearing, assemblages was found to be common. The close affinity of the sequences on Coll and Tiree to the type areas of Lewisian rocks in the extreme northwest of the Scottish mainland and in the Outer Hebrides was also noted by these workers.

The work of Drury and co - workers on the geochemistry of the meta - igneous rocks of Coll and Tiree has also shown them to be essentially similar to the rest of the Lewisian, especially the intensively investigated rocks of the Scourie - Laxford area and the Outer Hebrides ( see Weaver & Tarney 1980, 1981 and Tarney et al. 1979 in particular ). Drury ( 1974 ) divided the Coll and Tiree rocks into a suite of intermediate granulite - facies rocks with marked depletions in K, Th and Rb and distinctively high K/Rb, Sr/Rb and Ba/K ratios, and a more felsic suite of amphibolite - facies rocks with typical calc - alkaline igneous compositions. These latter were considered by Drury to be retrogressed granulite - facies gneisses which had undergone re - enrichment in K, Th and Rb. The degree of relative Rb depletion in the granulites is less than that found by Weaver & Tarney ( 1980 ) in rocks further to the north ( K/Rb c. 600, as compared to 600 - 2500 ). The K/Rb ratio is however still a factor of three lower in amphibolite - facies rocks in Coll and Tiree ( K/Rb  $\approx$  200 ) than in the granulites from the same islands. Drury's Th data for the granulites are at or below the detection limit for this element by the method used ( XRF ) but indicate Th depletion in the granulite - facies rocks, relative to the corresponding amphibolite - facies rocks, by a factor of at least 3: if the granulites are similar to those analysed by Weaver & Tarney ( 1980 ) Th depletions by factors of 10 to 50 would be expected. It seems likely, therefore, that the trace element ratios applied

elsewhere in the BTVP to distinguish granulite - facies Lewisian contaminants from amphibolite - facies Lewisian contaminants, principally K/Rb, Ba/Rb and Ba/Th ( e.g Thompson 1982 ) and Th/Ta and Th/Hf ( Thompson et al. 1980 ) can be applied validly in Ardnamurchan. Although isolated relict bodies of granulite - facies rocks occur in close association with amphibolite - facies rocks in Coll and Tiree, as in many other areas of the exposed Lewisian complex, the data of Smith & Bott ( 1975 ) and Bamford et al. ( 1977 ) indicate that on a large scale the Lewisian is composed of an upper amphibolite - facies layer and a lower granulite - facies layer. Bamford et al. ( 1978 ) propose a refinement of their earlier model, in which the granulite facies lower crust is divided into an intermediate - composition upper layer and a basic lower layer, immediately above the Moho. The boundary between the amphibolite and granulite facies rocks occurs at depths of between 6 and 14 kilometres ( Bamford et al. 1977 ) in the area studied by these workers. This lies well to the east and north of Ardnamurchan and well to the east of the Moine Thrust Zone ( section 2.2 ). The validity of extrapolating this crustal structure to the far side of the Moine Thrust Zone, over a distance of 150 - 250km, is placed in some doubt by the 'thick - skinned' tectonic model of the north - west Scottish Caledonides due to Soper & Barber ( 1982; see section 2.2 below ) which suggests that the crustal structure described by Bamford et al. lies above the Moine Thrust Zone at the latitude of Ardnamurchan and cannot be extrapolated through it to the west. However, the distinctive contamination patterns of the two main lava groups in Skye and the corresponding groups in Mull ( Thompson 1982; Dickin 1981 ) indicates that a similar large - scale division of the Lewisian exists in the Inner Hebrides as well, although the depth of the transition between the two types of Lewisian crust is not certain.

The relatively alumina - poor, calcareous and/or siliceous character of the Lewisian supracrustal metasediments on Coll and Tiree suggests that they are likely to have been mature sediments deposited in a high - energy, probably shallow - water environment in which clay minerals remained in suspension and heavy minerals were preferentially deposited. They can therefore be predicted to share the trace element characteristics of the Moinian and Jurassic sediments described below, such as low REE abundances and positive Hf and Zr anomalies caused by accumulation of zircon, although the relevant data is not available. However, these rocks are volumetrically minor and have relatively refractory bulk compositions and so would not be expected to be important contaminants. This argument applies for both bulk assimilation and selective assimilation of the most fusible lithologies, the two thermally plausible contamination scenarios discussed by Thompson et al. ( 1982 ) and

Dickin et al. ( 1984 ). In the former case the dominant potential contaminant present in Coll and Tiree is the intermediate 'grey gneiss', whilst in the latter it is the granitic gneisses and pegmatites.

## 2.2. Moine Group Metasediments.

Late Proterozoic metasediments of the Moine Group, with an age of at least 1100 Ma, although probably not much older ( Brook et al. 1977 ), form most of the outcrop in Northwest Scotland to the east of the outcrop of the Moine thrust ( Fig. 2.1 ). This thrust, which largely forms the western edge of the Caledonian orogenic belt in Scotland ( Coward 1983 ), is a moderately steeply eastward - dipping structure ( Powell 1974 ) at outcrop, although the interpretations of the geophysical evidence cited above ( see also Bott & Tantrigoda 1987 ) and most structural restorations of the fold belt ( Coward 1983 ) suggest that it shallows out to the east and underlies the Northwest Highlands at a depth of no more than 5 to 15 kilometres. However, Soper & Barber ( 1982 ) present an alternative interpretation of the structural and seismic data, in which they suggest that the Moine Thrust descends steeply to a depth of more than 20km beneath northwest Scotland and underlies granulite - facies rocks, probably including reworked Lewisian, which were metamorphosed to that grade during the **Caledonian** orogeny. If this model is correct then the Bamford et al. ( 1977 ) data cannot be used to infer the presence of a subhorizontal transition from amphibolite - facies middle crustal rocks to granulite - facies lower crust to the west of the Moine thrust except in the far north of Scotland. The Soper & Barber model has been criticised on the grounds that it is inconsistent with structural restorations of the imbricated Cambro - Ordovician rocks beneath the thrust in the far north - west of Scotland. These restorations suggest that the Moine Thrust does not cut down to lower crustal levels at distances of less than 60km east of its present - day outcrop ( Coward 1983 ). There are, however, no comparable constraints available from balanced sections through the southern part of the exposed Moine Thrust Zone, so the Soper & Barber model cannot be excluded for this section of the Moine Thrust Zone, which underlies Ardnamurchan and the area to the east.

Ardnamurchan lies toward the western edge of the outcrop of the Moine Group ( Fig. 2.1 ). The thrust itself does not outcrop south of Skye but its position can be constrained by:

(1) Outcrops of Moinian rocks as far west as Kilchoan, and the occurrence of cross - bedded quartzite

xenoliths, unlike any known Mesozoic rocks but very like the Moinian rocks, in cone sheets at 46077041, on the north coast of Ardnamurchan.

(2) The occurrence on Coll and Tiree of Lewisian rocks lacking Caledonian deformation and hence forming part of the Caledonian foreland.

(3) The lack of intense, mylonitic deformation of the Moinian rocks in Ardnamurchan, such as occurs for some hundreds of metres above the Moine thrust in most areas where it does outcrop.

(4) Extrapolation along the strike of the thrust belt from Skye and the western limit of Moinian rocks in south - west Mull ( although the presence of the Strathconon fault introduces a considerable degree of uncertainty into the extrapolation ).

These constraints suggest that the Moine Thrust Zone probably underlies Ardnamurchan at a depth of a few kilometres, at least as far east as the Sgurr Bheag Slide ( Fig. 2.1 ). The close proximity of outcrops of Lewisian and Moinian rocks on either side of the inferred subcrop of the thrust between Ardnamurchan and Coll, and between Mull and Iona to the south, suggests that there can be no great thickness of sediments of the Torridonian or Cambro - Ordovician sequences beneath the thrust and that for present purposes it can be considered to juxtapose Lewisian rocks of the Caledonian foreland and the rocks of the Caledonian fold belt.

In principle, therefore, it should be possible to distinguish, by the geochemistry of the contaminants, contamination of Tertiary magmas taking place below the Moine thrust from that taking place above it, provided that the volume of Lewisian ( or Grenvillian ? ) meta - igneous rocks incorporated into the Caledonian belt and transported as far west as Ardnamurchan by movement on the Moine thrust is not significant. Inliers of Lewisian rocks outcrop within the Moine sequence north and east of Ardnamurchan, in Morar and Moidart, but only to the east of an area of westward - directed thrusts in the steep to inverted western limb of a major north - south antiform, the Morar anticline of Johnstone et al. ( 1969 ). It should be noted that the more extensive Lewisian and Grenvillian inliers of Skye, Glenelg and Torridon lie north of the Strathconon fault and cannot be simply projected along strike: since this fault is probably sinistral the subsurface extensions of these inliers to the south of the fault would lie further to the east than a simple along - strike projection would suggest.

Furthermore, Johnstone et al. ( 1969 ) consider that the Moinian rocks in Ardnamurchan belong to the uppermost part of the Morar Division of the Moinian succession. The thickness of this unit was estimated as 7 kilometres by J.V.Watson ( 1963 ) and although Johnstone et al. suggest that this may be an overestimate because of repetition of parts of the sequence by thrusting, this does indicate that there is very little space within which older rocks could occur between the base of the Moine succession and the position of the Moine thrust zone in this area, as predicted by both Coward ( 1983 ) and Soper & Barber ( 1982 ). Even if pre - Moinian rocks are present above the Moine thrust zone beneath Ardnamurchan, they must be very close to it, in which case the crustal sequence shown in Fig. 2.9 below will remain valid for the purposes of modelling the siting of crustal contamination.

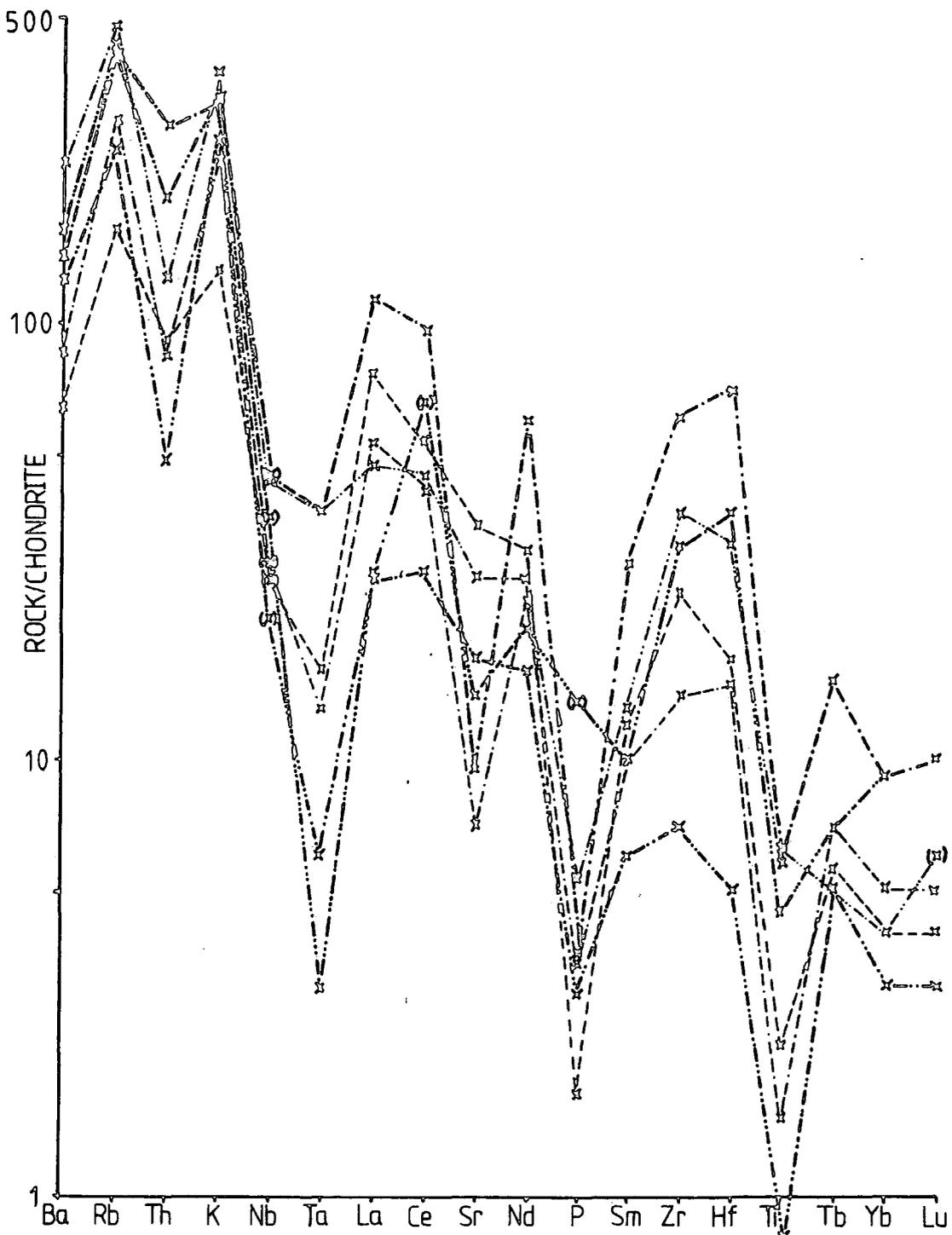
According to Johnstone et al. ( 1969 ) the Morar division consists of a thin basal pelitic unit and two dominantly psammitic units, the Lower and Upper Morar Psammities, which are separated by a second relatively thin pelitic unit, the Morar Schists. Impure calc - silicate bands were considered by Johnstone et al. to be a ubiquitous minor component of the psammities but, although carbonate, clinozoisite and/or epidote and actinolitic hornblende are common minor components of the psammities collected in the course of this work, distinct calc - silicate rocks were not found. Richey et al. ( 1930 ), in the course of their work in western Ardnamurchan, divided the Moine rocks in the area into a western group, the 'Kilchoan Moines', and an eastern group characterised by higher grade metamorphism and more intense deformation. The 'Kilchoan Moines' are characterised by low grade greenschist facies assemblages ( quartz + detrital microcline + detrital plagioclase + albite + clinozoisite + chlorite  $\pm$  muscovite  $\pm$  carbonate ) and shallowly dipping, only weakly folded bedding, whilst to the east of a NNW - SSE line running through Kilmory and Ardslignish, at the eastern edge of the Tertiary complex ( see Fig. 1.2 ), there is a rapid transition to higher grade metamorphic assemblages with biotite, garnet and/or hornblende and fewer recognisable relict detrital grains. This transition coincides roughly with the western edge of a belt of steep NNW - SSE trending tight folds, but the work of Johnstone et al. implies that the rocks outcropping to the east should underlie the 'Kilchoan Moines'. Both groups of rocks consist of interbedded psammite - semipelite sequences, although the proportions of different lithologies may vary between the two. Samples from both were collected for geochemical analysis, but any variation between them is insignificant compared with variation between lithologies within them ( see below ).

The outcrops of Moine series rocks examined are dominated by decimetre to metre thick, laterally continuous beds of meta - sandstone, with some pebbly units, separated by much thinner micaceous schistose semi - pelites. Cross - bedding is common within the meta - sandstone. The overall sedimentary environment was probably fluvial, of braided stream type, or shallow marine ( R. Glendinning pers. comm. ). In any case it was certainly a high - energy environment, with little opportunity for mud or limestone deposition. Johnstone et al. ( 1969 ) note the presence of detrital heavy mineral accumulations, particularly in the Lower Morar Psammites, and although these are not as obvious in the rocks exposed in western Ardnamurchan there are a number of features of the bulk geochemistry of these rocks which can be interpreted in terms of their depositional environment.

Analyses of these rocks ( see Appendix 2 ) are characterised by relatively high  $\text{SiO}_2$ , and low abundances of all the constituents of clay minerals, except for K and Rb, which are anomalously high because of the presence of quite large amounts of detrital microcline in the rocks ( see Fig. 2.2; the Rb positive anomaly may be an analytical artefact ( see Appendix 1 ) but the K anomaly is a real feature of the rocks ). The presence of small amounts of carbonate in the original sediments means that the rocks are not strongly peraluminous on the whole. Their most characteristic geochemical feature, however, is generally high but very variable Zr and Hf relative to middle and heavy rare earth elements. This can be quantified as the chondrite - normalised ratio  $\text{Hf}_N/\text{Sm}_N$  ( preferred here to  $\text{Zr}/\text{Sm}$  because the Hf determinations made were in general more reliable than those for Zr ( see Appendix 1 )) which ranges from about 1 to 3.7 in the rocks analysed. This contrasts with values of these ratios in most igneous rocks, except those with a sedimentary source rock component and extremely fractionated rocks, which are close to unity because of the similar distribution coefficients of Hf, Zr and MREE between melts and most cumulate mineral assemblages which contain more than trace amounts of them ( see Henderson 1982; Dunn 1987; E.B.Watson et al. 1987 ), except of course for highly fractionated assemblages which contain minerals such as zircon and allanite.

The most plausible explanation of the generally high but varied Hf/Sm ( and Zr/Sm ) ratios in these sediments is that they reflect preferential accumulation of detrital zircon in sediments deposited under high - energy conditions: an analysis of a Moinian pelite from S.W. Mull ( Thompson et al. 1986 ) shows a complementary depletion in Hf and Zr relative to MREE although absolute abundances ( 4ppm Hf and 147ppm Zr ) are within the range of values in the psammites ( 1-14ppm Hf, 51-420ppm Zr ). A similar pattern is shown by Jurassic rocks in the area ( section 2.3,

2.2 Chondrite - normalised minor and trace element data for Moirian Rocks.



- |             |       |       |       |
|-------------|-------|-------|-------|
| -----       | 157/1 | ----- | 159/1 |
| - . - . - . | 157/2 | ----- | 173/1 |
| .....       | 158   | ----- | 173/2 |

} From "Kilchoan Moines" of Richey et al. 1930

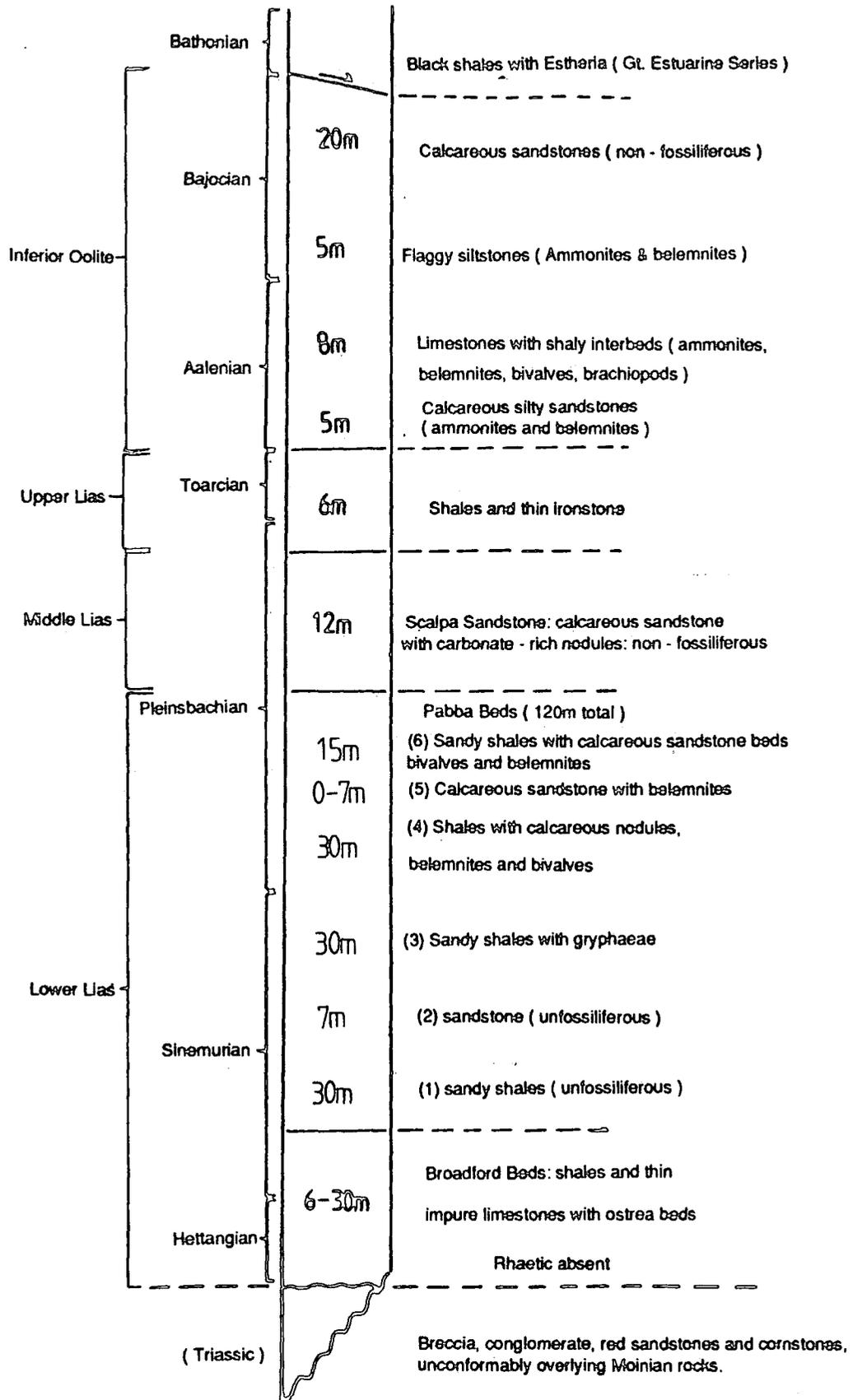
below ). In contrast to the generally high but scattered values in the Moinian sediments, values of  $Hf_N/Sm_N$  in Lewisian rocks are in the range 0.5 to 1.5, except for the various highly fractionated trondhjemites and pegmatites (  $Hf_N/Sm_N$  up to 4 ). The latter can however be easily distinguished from the Moinian rocks by their very much higher chondrite - normalised LREE / HREE ratios ( see below ). Strictly, therefore, contamination by Moinian or similar sedimentary rocks would be expected to produce a wide scatter in Hf/Sm and Zr/Sm ratios, depending on where in the sequence defined by Johnstone et al. contamination took place. The semi - pelites within the Upper Morar Psammite analysed during the course of this work have positive Hf and Zr anomalies as great as, or even larger than, those of the psammites, so contamination taking place anywhere in this particular unit would produce high Hf/Sm and Zr/Sm ratios in the contaminated rocks. Structural work by Johnstone et al. suggests that this unit should form most of the Moinian sequence beneath westernmost Ardnamurchan, so it seems likely that in fact any Moinian contaminants present in rocks of the Hypersthene Gabbro would have high average Hf/Sm.

Other features of the Moinian rocks analysed for the elements concerned ( see Appendix 2 ) are high Sr(ppm)/CaO(Wt%) ( 100 to 3000 ), high Rb/Ba ( 0.05 to 0.1 ), very low  $P_2O_5$  and fairly uniform  $La_N/Yb_N$  , around 10 to 20. This latter is very uniform compared to values from Lewisian rocks, which are also much higher in the case of the more evolved Lewisian trondhjemites and pegmatites ( 50 to 500 ( Weaver & Tarney 1980,1981 )).

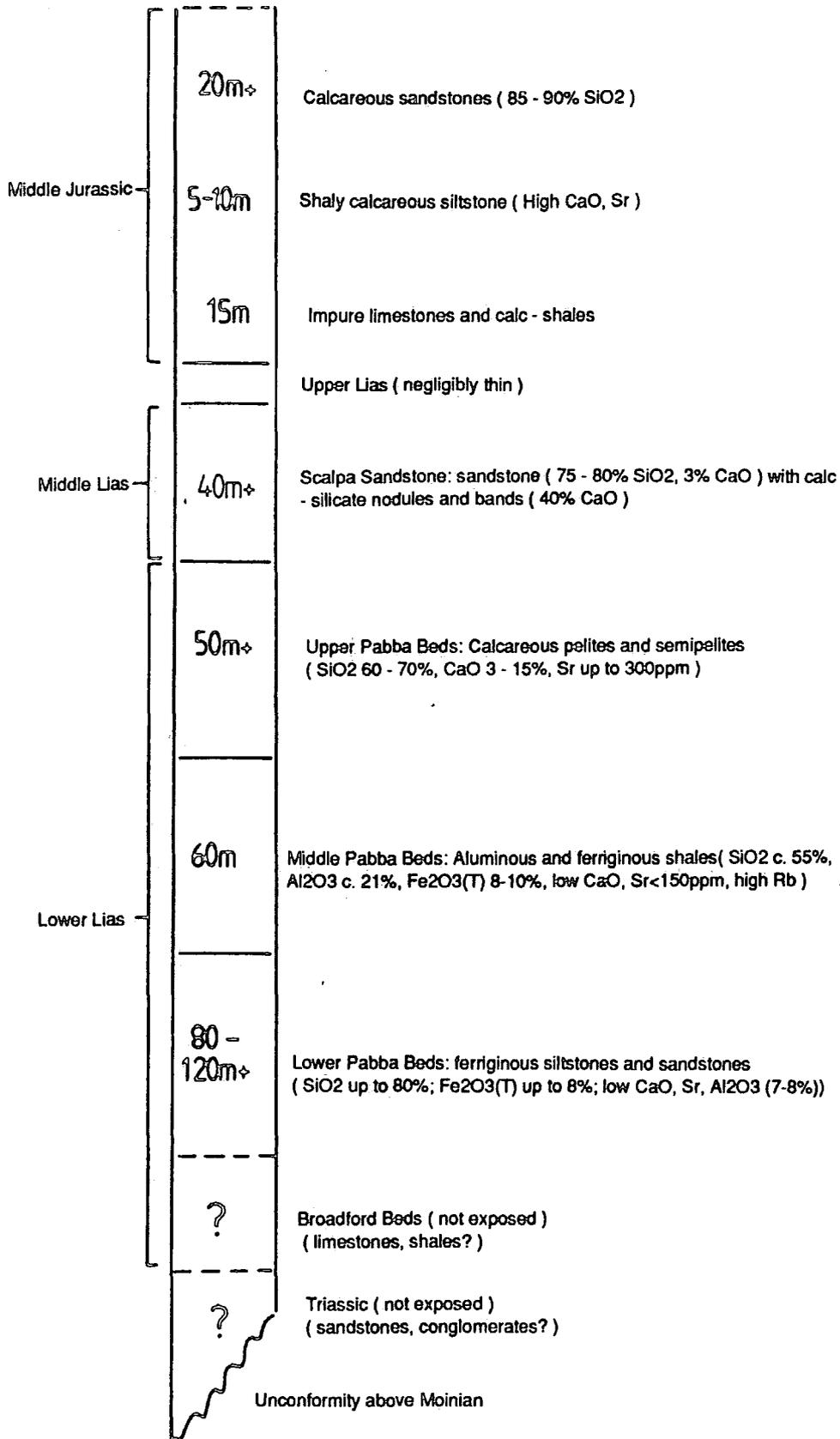
### 2.3 Mesozoic Sediments.

The Mesozoic rocks of Ardnamurchan, which are all Jurassic in age apart from a thin Triassic unit at the base of the succession, were described in detail by Richey et al. ( 1930 ), following on from earlier work by Judd ( 1874 ) in particular. Despite the Tertiary metamorphism encountered in most outcrops, even in the eastern part of the complex and around Kilchoan Bay, where most of their work was done, these workers were able to construct a fairly complete biostratigraphy and correlate the succession in Ardnamurchan with those in other parts of the series of Mesozoic basins in northwest Scotland ( Fig. 2.1; see Richey 1961; Anderton et al. 1979; Hudson 1983 for general reviews of the Mesozoic geology of northwest Scotland ). Many of Richey et al.'s units have distinctive and characteristic lithologies ( Fig. 2.3 ), so it was possible to map them by lithology during the course

2.3 Chronostratigraphy of the Mesozoic of Ardnamurchan, modified after Richey et al. ( 1930 ) and Anderton et al. ( 1979 ) ; included here for comparison with Fig. 2.4.



2.4 Informal lithostratigraphy used in this work for correlation of Mesozoic rocks west of the Glas Eilean fault.



of the present work. However, substantial differences were found between Richey et al.'s descriptions of the Lower Liassic rocks, which were largely based on exposures well to the east of the Hypersthene Gabbro, and the Lower Lias on the southern side of the pluton. These differences, which involve both lithological variation and a westward thickness increase, indicate the presence of Mesozoic structures in the area which are dealt with in section 2.5. These appear to form the eastern margin of the lower Mesozoic basin in this area and are also reflected in the position of the late Mesozoic unconformity in the area, which cuts down through the Mesozoic succession to lie directly on Moinian rocks in the extreme south - east of the Tertiary complex.

### **2.3.1 Triassic and Lower Jurassic.**

The oldest Mesozoic rocks in Ardnamurchan, conglomerates, sandstones and cornstones of probable Triassic age and the limestones and calcareous shales of the Broadford Beds ( Fig. 2.4, above ) are only exposed to the east of a large fault, here named the Glas Eilean fault after the island which forms its southernmost outcrop at 483627. This fault runs north from Glas Eilean to the edge of the Tertiary complex at about 485647 ( see Map 1 ) and continues north - north - west across Centre 3 of the Tertiary complex as a series of ductile shear zones, crush zones and distinct faults to reappear on its northern side as a major fault which reaches the coast just east of Rubha Carrach. Younger Jurassic rocks outcrop to the west of this fault so it is reasonable to suppose that these Triassic and lowest Jurassic rocks may be present beneath the surface around the Hypersthene Gabbro. Richey et al. ( 1930 ) suggested that a hornfelsed calcareous unit found on the summit of Dubh Chreag beneath pelites of the Pabba Beds might be Broadford Beds limestone, but this was not found during the course of the present work and if present would overlie the Lower Pabba Beds as defined below ( nodular calc - silicate bands are present in the Middle Pabba Beds and it may have been an example of one of these that was found ). However, the Upper Broadford Beds in south Skye are silty shales and ferruginous siltstones and some of the ferruginous sandstones and siltstones informally assigned to the Lower Pabba Beds, below, may be lateral equivalents of these and of the limestones in eastern Ardnamurchan. Work by Hudson ( 1964 ) and Steel ( 1977 ) suggests that the sediment sources for the northern half of the Inner Hebrides basin, which includes Ardnamurchan, lay to the north and west. This would allow the deposition of clastic sediments in westernmost Ardnamurchan whilst deposition of limestones continued to the east.

Shales, ferruginous siltstones and rare sandstones and limestones assigned to the Pabba beds by Richey et al. ( 1930 ) form the thickest unit of the Mesozoic in Ardnamurchan. They form most of the exposed Mesozoic succession around the southern and south - western sides of the Hypersthene Gabbro, and also occur in smaller areas at Glendrian Bay and near Duin Bhain ( see Map 1 ). On the southern side of the pluton at least the Pabba Beds show a three - fold division into distinct units. This division is an informal one and only applies to the outcrops investigated, which lie west of the Glas Eilean fault. The fossiliferous units ( 3 - 6; see Fig. 2.3 ) of the Pabba beds defined by Richey et al. ( 1930 ) are considered here to form the Upper Pabba Beds; the sandstone unit (2) is not well - developed in any of the sequences examined although thin fine - grained sandstones do occur near the boundary of the Middle and Upper Pabba beds as defined below; whilst Richey et al.'s unit (1) corresponds to the Middle Pabba Beds as defined here and the 'Lower Pabba Beds' defined below may in fact be a lateral equivalent of the Broadford Beds limestones to the east of the Glas Eilean fault, although palaeontological evidence is lacking. The question of whether or not this correlation is correct is not critical to the present work except as an indication that the Glas Eilean fault may have been active as early as the lower Jurassic.

The oldest unit was originally a thick sequence of plane - bedded ferruginous siltstones and fine - grained sandstones with thin pelitic interbeds. It is only exposed close to the contact with the Hypersthene Gabbro and has been intensely hornfelsed ( see Chapters 3 and 5 ). The thickest exposed sequence of these Lower Pabba Beds occurs around 447630, south and west of Hill 210 on the southern margin of the pluton. After an approximate correction has been made for the effects of cone sheets on the apparent thickness of the unit, its thickness can be estimated to be at least 80 to 120 metres, the base of the sequence being beneath the surface. This is almost as much as the entire thickness of the Lower Lias in Ardnamurchan according to Richey et al. ( 1930 ) and indicates substantial thickening of the sequence to the west of the Glas Eilean fault.

The Lower Pabba beds pass up, without any obvious breaks, via as much as ten metres of grey, fine - grained quartzose meta - siltstones, into the Middle Pabba Beds. The Middle Pabba Beds are very distinctive, finely laminated pelites, with alternating dark very fine - grained pelitic and slightly paler, more quartzose laminae. They were highly micaceous prior to metamorphism but also contain nodular quartz - clinopyroxene nodules which were probably originally ferri - calcareous concretions ( sideritic or ankeritic ). These Middle Pabba Beds are as much as 60 metres thick, although the

top, like the base, is gradational and is difficult to locate precisely.

The Upper Pabba Beds are rather more varied than the two older units, and include siltstones, calcareous and micaceous sandstones and impure limestones in addition to the slightly calcareous pelites which form the bulk of the unit. Unlike the two older units, they contain a moderately abundant although poorly preserved fossil assemblage. Gryphaeae and belemnites are the commonest recognisable fossils and persist to relatively high grades of metamorphism, being the thickest - shelled part of the assemblage. Fragmentary shell debris is very common, and is a useful identification feature in the field.

The sequence described above is only fully developed in the area between Kilchoan and An Acairseid. Part of the same sequence is recognisable on Druim na Cloise ( see Map 1 ), where a small thickness of interbedded ferruginous meta - sandstones and pelites, equivalent to the Lower Pabba Beds described above, outcrops at the base of a Middle Pabba Beds sequence which is as much as 80m thick. Finally, a few intensely disrupted areas of Pabba Beds lithologies occur on the northern margin of the pluton, most notably on the south - east side of Glendrian Bay ( Grid Sq. 4670; see Map 1 and section 3.2.2 ). These outcrops are intensely faulted and intruded by a great number of cone sheets, felsitic sheets and rheomorphic breccia bodies but appear to underlie Scalpa Sandstone ( Middle Liassic ) which outcrops to the north - east. They are composed of ferruginous and aluminous pelites ( Middle Pabba Beds ? ) and overlying calc - pelites and calc - semi - pelites with calc - silicate nodules ( Upper Pabba Beds ? ). The thickness of this succession is uncertain, because of the deformation, but is at least a few tens of metres ( see Map 2 ).

The three units of the Pabba beds are compositionally very distinctive. The Lower Pabba Beds are silica - rich (  $\text{SiO}_2$  up to 80wt% ), have high total Fe (5-8% as  $\text{Fe}_2\text{O}_3$ ) and Fe/Al ( hence the dominance of hypersthene over cordierite in the metamorphosed rocks ), low  $\text{Al}_2\text{O}_3$  (7-8% ) and CaO ( c. 0.7% ). They commonly contain metamorphic hornblende as well as cordierite at moderate grades of metamorphism. The Middle Pabba Beds are very uniform, being ferruginous pelites with about 55%  $\text{SiO}_2$ , 21%  $\text{Al}_2\text{O}_3$ , high  $\text{Fe}_2\text{O}_3^T$  ( c. 8% ), high  $\text{K}_2\text{O}$  but up to 3% CaO. Finally, the Upper Pabba Beds are more heterogenous, but tend to have markedly higher Si and Ca, and lower Al, Fe and K, than the Middle Pabba Beds. The trace element characteristics of the three units are also distinctive in many respects, the principal differences being

- (1) REE element abundances ( highest in the pelites of the Middle Pabba Beds ).  $La_N/Yb_N$ , however, is relatively constant at 8 - 10.
- (2) Rb and Rb/Ba( again, highest in the Middle Pabba Beds ).
- (3) Cr ( up to 150ppm in the Middle Pabba Beds, and less than 100ppm in the other units ).
- (4)  $Hf_N/Sm_N$  is less than 1 in the Middle Pabba Beds, reflecting their low - energy environment of deposition and consequent pelitic character ( see Fig. 2.5, below, in which a strong overall correlation between Hf/Sm and  $SiO_2$  is apparent for all the Jurassic rocks, reflecting the preferential accumulation of quartz as well as zircon in high - energy environments of clastic sedimentation, although in detail the two elements do not always correlate ( see section 4.2.4 )).
- (5) higher Sr in the Middle and Upper Pabba Beds ( 200 - 300ppm, compared to 140 - 150ppm in the Lower Pabba beds ).

The Middle Lias in Ardnamurchan, as in many other parts of the Inner Hebrides, is represented by the quartzites and calcareous sandstones of the Scalpa Sandstone. Richey et al. ( 1930 ) describe a complete section through this unit on the north coast of the peninsula, about 1km east of Rubha Groulin ( Grid Ref. 478711 ). This sequence, which lies to the east of the Glas Eilean fault, is only 10 - 15m thick. The main exposures west of this fault, to the west of Maol Bhuidhe on the south side of the peninsula ( around 454627 ) and at Glendrian Bay on the north coast, are about 40m and at least 40m thick respectively. The Scalpa Sandstone appears, therefore, to show sharp lateral thickness variation across the fault, particularly in the case of the two occurrences on the north coast which are less than 2km apart. In the field it is a pale yellowish to white sandstone with little internal structure apart from calcareous nodules and nodular calcareous beds: these are present throughout the Scalpa Sandstone on the south coast but die out towards the faulted top of the sequence at Glendrian Bay. Compositionally, the sandstones in the unit are characterised by high  $SiO_2$  ( 75 - 80% ) and only moderate CaO ( c. 3% ): a calc - silicate rock within meta - sandstones that form part of the Scalpa Sandstone ( see section 3.2.3 ) and outcrop close to the contact east of Sanna Point ( Grid Ref. 445703 ) has 40% CaO, 52%  $SiO_2$  and little else.

In contrast to the unusually thick Lower and Middle Liassic sequences in western Ardnamurchan, the Upper Lias is very thin, being represented by 6–7m of dark organic - rich shales ( now metamorphosed to black sulphide - rich pelite in its occurrences in western Ardnamurchan ) and a thin meta - ironstone. Apart from the latter the Upper Lias is likely to be compositionally similar to the Middle Pabba Beds but in any case is volumetrically insignificant.

### 2.3.2 Middle Jurassic .

In situ Middle Jurassic rocks in Ardnamurchan are only exposed in downfaulted blocks on the southern side of the Hypersthene Gabbro, although Richey et al. ( 1930 ) recorded blocks of sandstone containing Middle Jurassic fossils in Tertiary agglomerates on the eastern side of the Ardnamurchan complex ( Grid Ref. 520720 approx. ). At the time, since it was believed that all the agglomerates were vent - infills this was taken to indicate the presence of Middle Jurassic rocks beneath the surface in that area. However, more recent re - interpretations of the agglomerates ( section 2.4 ) cast doubt on this: the youngest Jurassic rocks definitely present in that area are Liassic in age. The main exposures of Middle Jurassic rocks are in a downfaulted block centred on Maol Bhuidhe ( see Map 1 and Fig. 3.17 ) but rocks of the same sequence also occur in elongate downfaulted blocks immediately adjacent to the southern margin of the Hypersthene Gabbro itself. Richey et al. describe the following sequence of Middle Jurassic ( Aalenian - Bajocian ) rocks from Maol Bhuidhe:

- 20m calcareous sandstone
- 5m calcareous shaly siltstone
- 8m impure limestone with shaly bands
- 5m calcareous sandstone

During the present work, the lower two units were found to grade into each other and are perhaps best considered as a single unit grading up from a sandstone into a limestone: they are mapped as such on Fig. 3.17. The shaly siltstone unit also appears to be rather thicker than was indicated by Richey et al., although this may be a consequence of misidentification of the lower parts of the sandstone unit at the top of the Bajocian succession, which are more silty than the massive calcareous sandstones at the top of the unit. Analyses of these latter ( 243 and 246, Appendix 2 ) show them to be similar to the Scalpa Sandstone, but with an even higher SiO<sub>2</sub> content ( over 90%

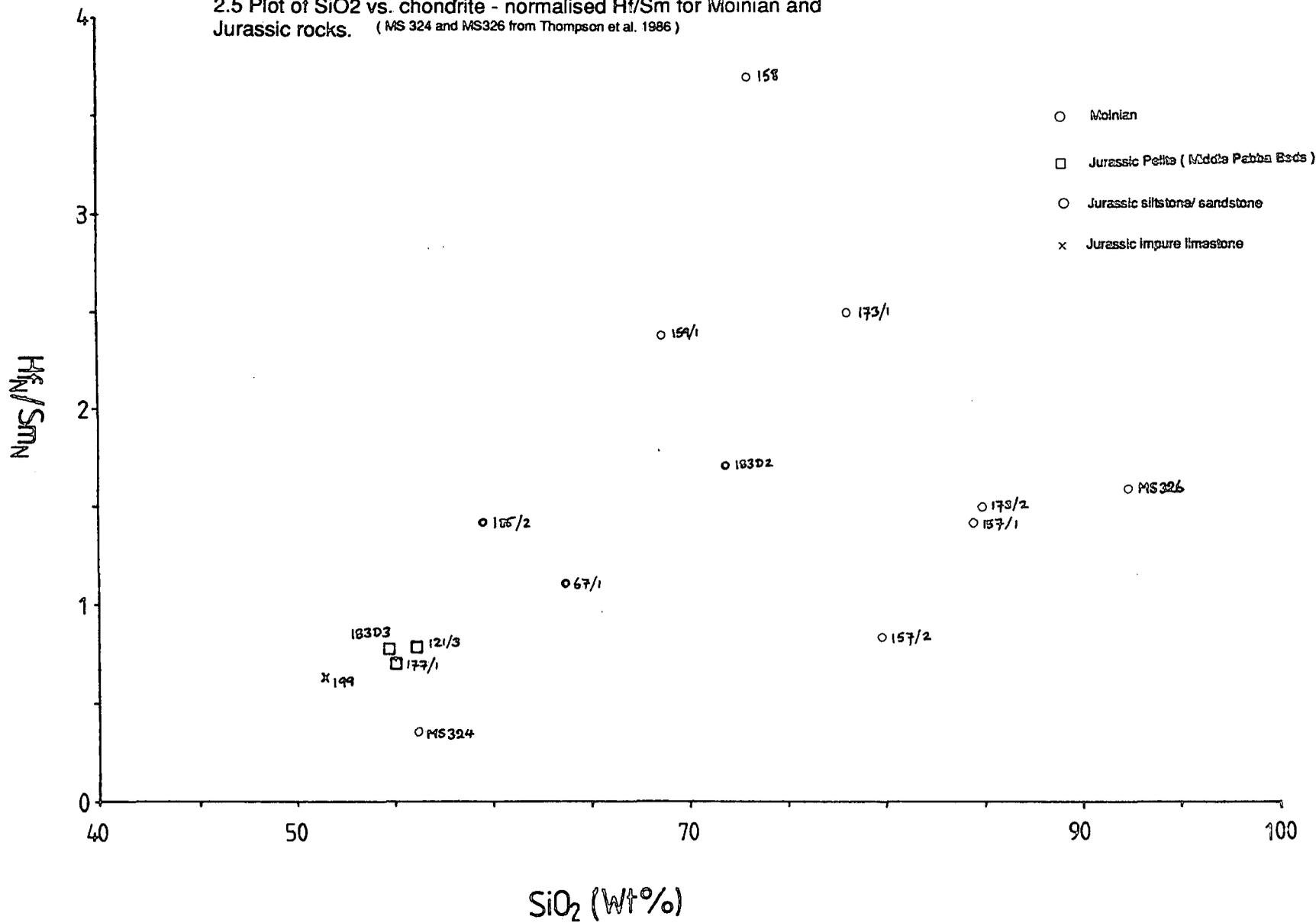
). The silty beds are both richer in CaO and in alumina and alkalis. Apart from a small block of Bathonian ( Richey et al. 1930 ) shale in a Tertiary vent at Cuingleum, on the southeast side of Maol Bhuidhe, these are the youngest Mesozoic rocks in Ardnamurchan.

The same units can be recognised, despite high - grade contact metamorphism, in the downfaulted blocks which run along the southern contact of the Hypersthene Gabbro. These have been intensely disrupted by the faulting ( see sections 3.2.5.1 and 3.2.5.4 ) with the result that complete successions are not identifiable. The dominant rock type at outcrop is a calcareous semipelite, analysed samples of which ( 67/1 and 81/3 ) have high CaO and also very high Sr ( 615 and 471 ppm respectively ) which results in a  $Sr(ppm)/CaO(wt\%)$  ratio of about 50, a value more typical of the Jurassic pelitic rocks than of other originally carbonate - rich rocks in the Jurassic succession. These rocks correspond lithologically to the calcareous siltstones in the succession at Maol Bhuidhe, and like them, are underlain by calc - silicates : this outcrops at about 46786384 in the form of an idocrase - grossular rock. At the western end of Druim na Gearr Leacainn, weakly bedded quartzite hornfelses are present which probably correspond to the less massive, slightly silty sandstones at the base of the thick sandstone unit which forms the bulk of the sequence at Maol Bhuidhe ( see above ).

### **2.3.3. Summary of the Mesozoic sequence and its geochemical characteristics.**

The Mesozoic succession west of the Glas Eilean fault is summarised in Fig. 2.4, together with the geochemical characteristics of the various units. As far as can be seen, this sequence was almost uniform around the Hypersthene Gabbro prior to the deformation associated with it and can therefore be used to determine fault offsets and relative uplifts associated with it. The sequence to the east of the fault, however, is much thinner and was also eroded much more deeply prior to the eruption of the Tertiary lavas. This erosion was especially severe in the south - east of the Tertiary complex, where the sub - basalt unconformity cuts down through the Liassic and Triassic sediments into the Moine Group metasediments. In contrast, rocks as young as the top of the Upper Liassic outcrop on the north coast to the east of the fault, between Rubha Groulin and Faskadale. This suggests that the area may have acquired a very slight northward tilt prior to the formation of the sub - basalt unconformity, between the Middle Jurassic and the very earliest part of the Tertiary. The geometry of this unconformity also constrains the Glas Eilean fault, or a fault very close to it, to have had a pre - Tertiary history of movement with downthrow to the west: the greater thickness of

2.5 Plot of SiO<sub>2</sub> vs. chondrite - normalised Hf/Sm for Moinian and Jurassic rocks. (MS 324 and MS326 from Thompson et al. 1986)



the Jurassic succession and of units within it to the west of the fault indicates that it was probably active as a more - or - less synsedimentary fault.

Geochemically, the Jurassic sequence is very variable and shows considerable compositional overlap with the Moinian sediments. Amongst major elements, the higher Fe and Ca contents of most of the Jurassic rocks are the most distinctive properties of the group. With regard to minor and trace elements, chromium, vanadium, strontium and phosphorus are higher in the Jurassic rocks by factors of 2 or more ( see Appendix 2 ). Unfortunately these elements are strongly affected by fractional crystallisation in basic to intermediate magmas and their usefulness as indicators of whether contamination of a magma has involved Moinian or Jurassic rocks ( and hence of the site of contamination ) depends on an understanding of the contamination processes involved and whether they are linked to fractional crystallisation ( see Chapter 4 ). As is the case in the Moinian sediments, the processes of sedimentation have resulted in accumulation of detrital zircon in those sediments deposited in higher - energy sedimentary environments, leading to the strong correlation between Hf/Sm ratio and silica content of the Jurassic sediments shown in Fig. 2.5.

## **2.4. Pre - Hypersthene Gabbro Tertiary Rocks.**

As noted in Chapter 1, the Hypersthene Gabbro was considered by Richey et al. ( 1930 ) to be the earliest large pluton exposed at the surface in Ardnamurchan but also to be predated by a wide variety of igneous rocks, together with a few thin sedimentary units. These can be divided into three main groups: basalts and picritic basalts of plateau basalt type ( with associated sills, dykes and thin sediments ); volcanoclastic rocks belonging to Centre 1 of Richey et al.s' subdivision of the complex; and various small gabbroic to granitic intrusions. Most of the latter also belong to Centre 1.

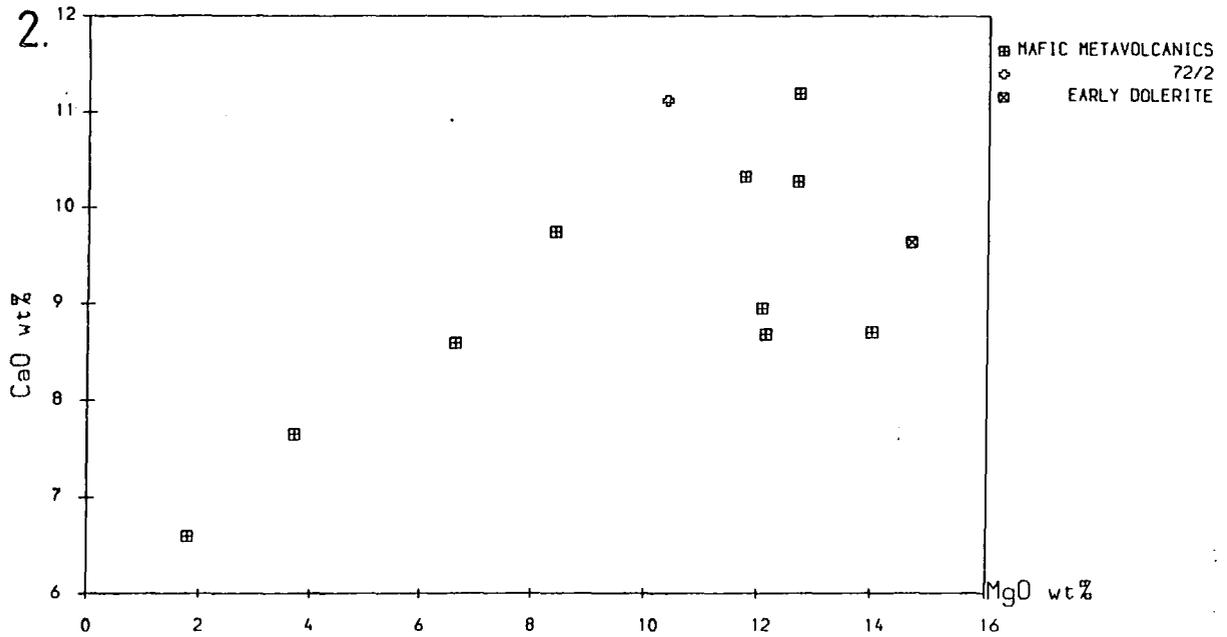
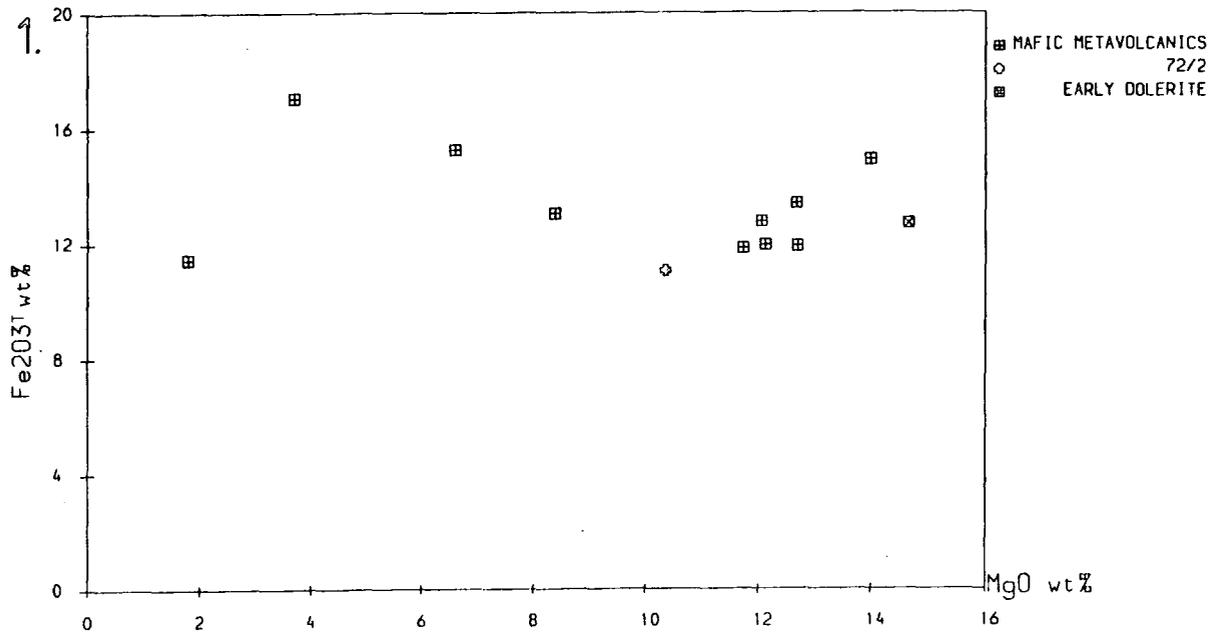
### **2.4.1. The Plateau Basalt Sequence.**

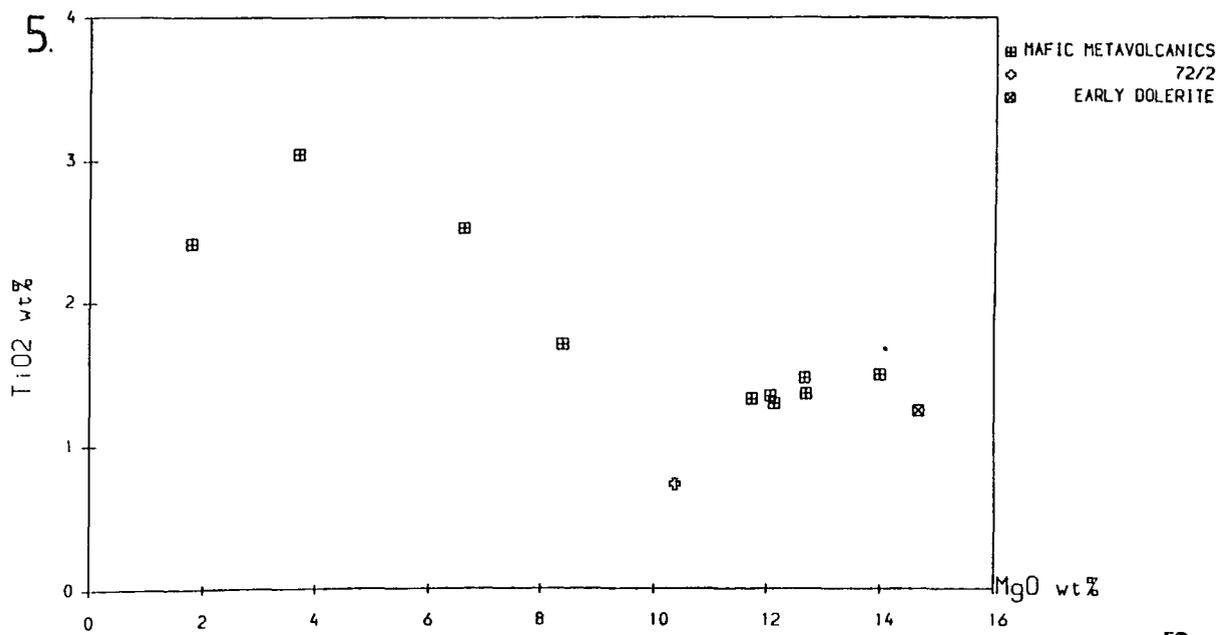
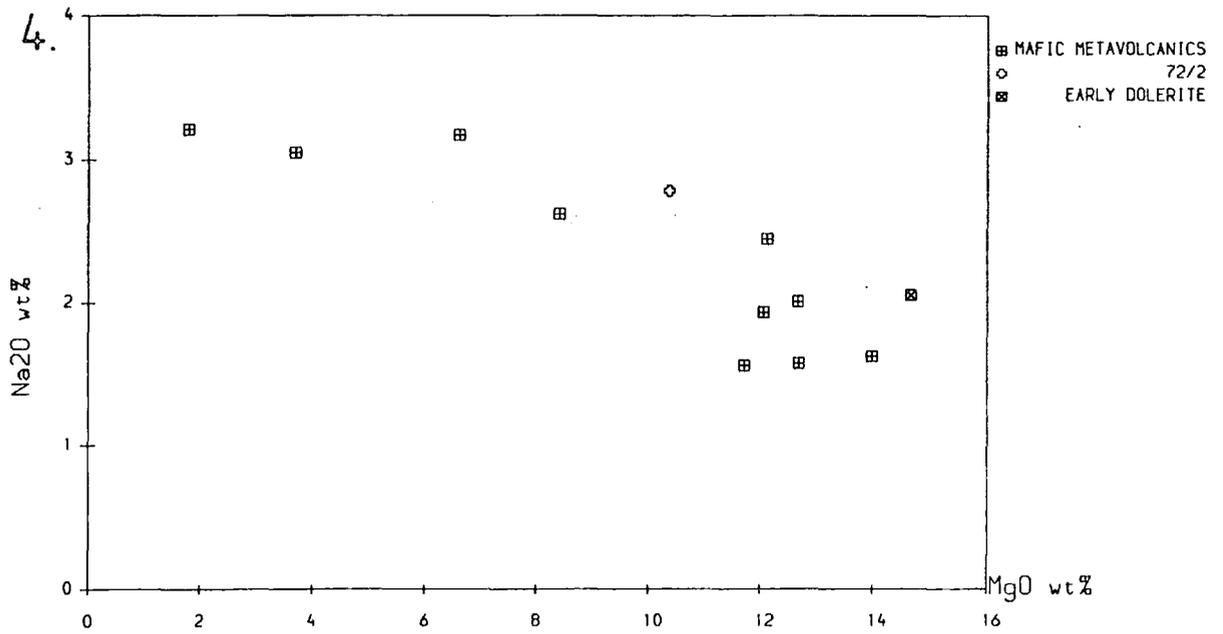
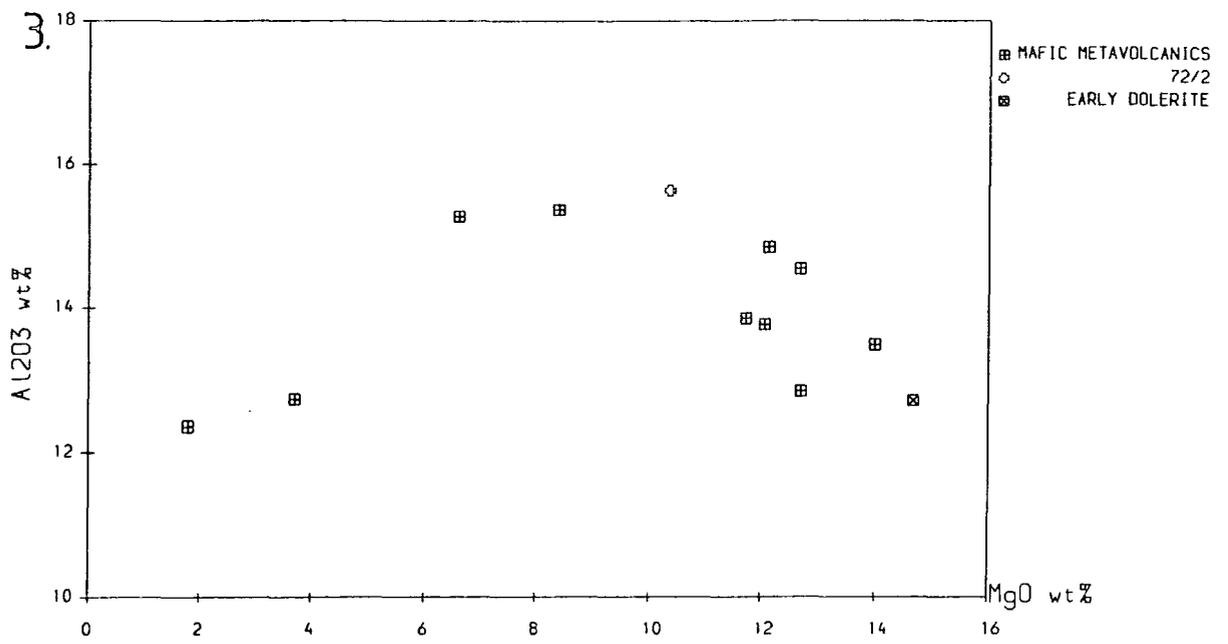
As is the case in all the Tertiary volcanic areas in north - west Scotland, the oldest Tertiary igneous rocks in Ardnamurchan are a suite of transitional, mildly alkaline to mildly tholeiitic, basaltic rocks characterised by a high content of incompatible elements, most notably Ti ( see below ), for their

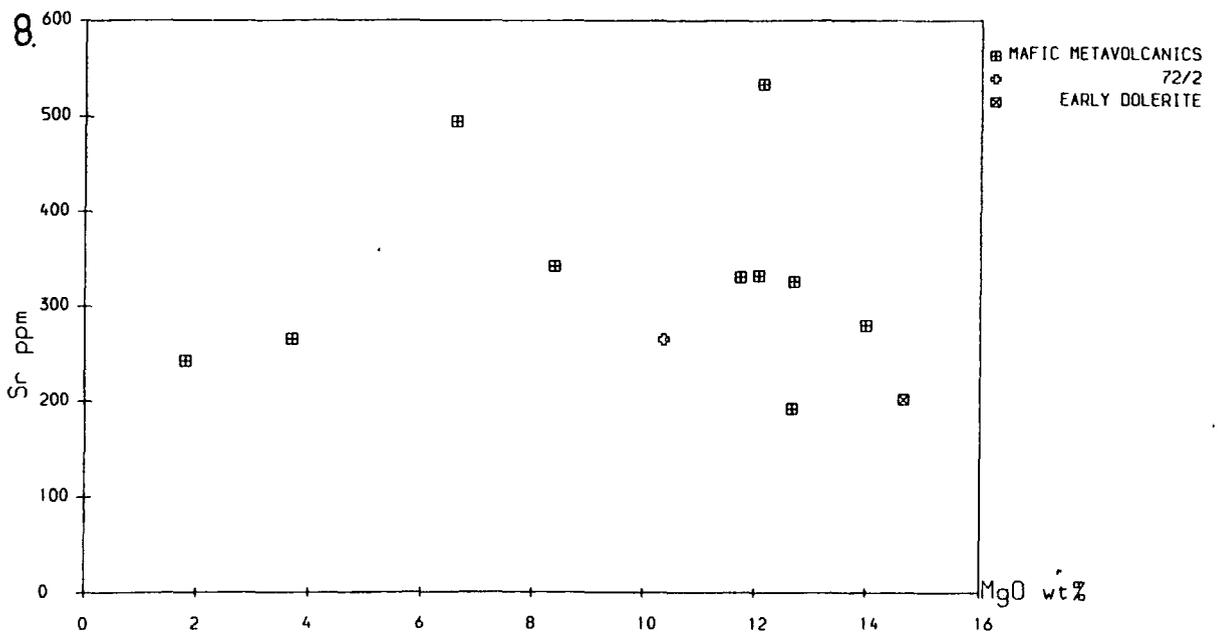
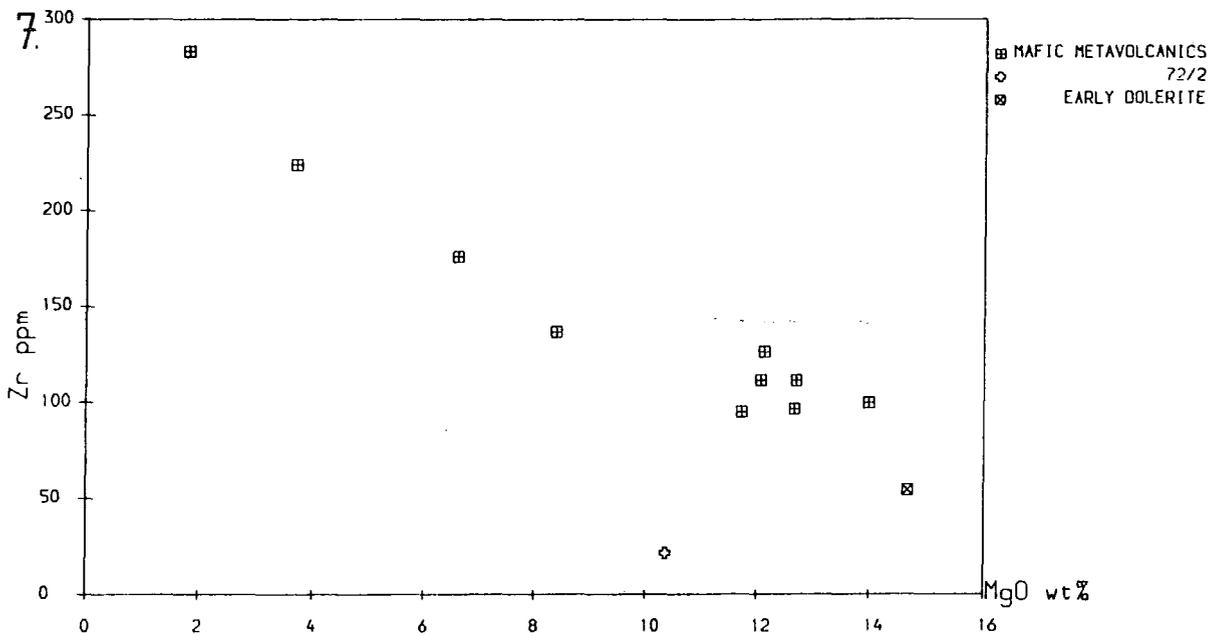
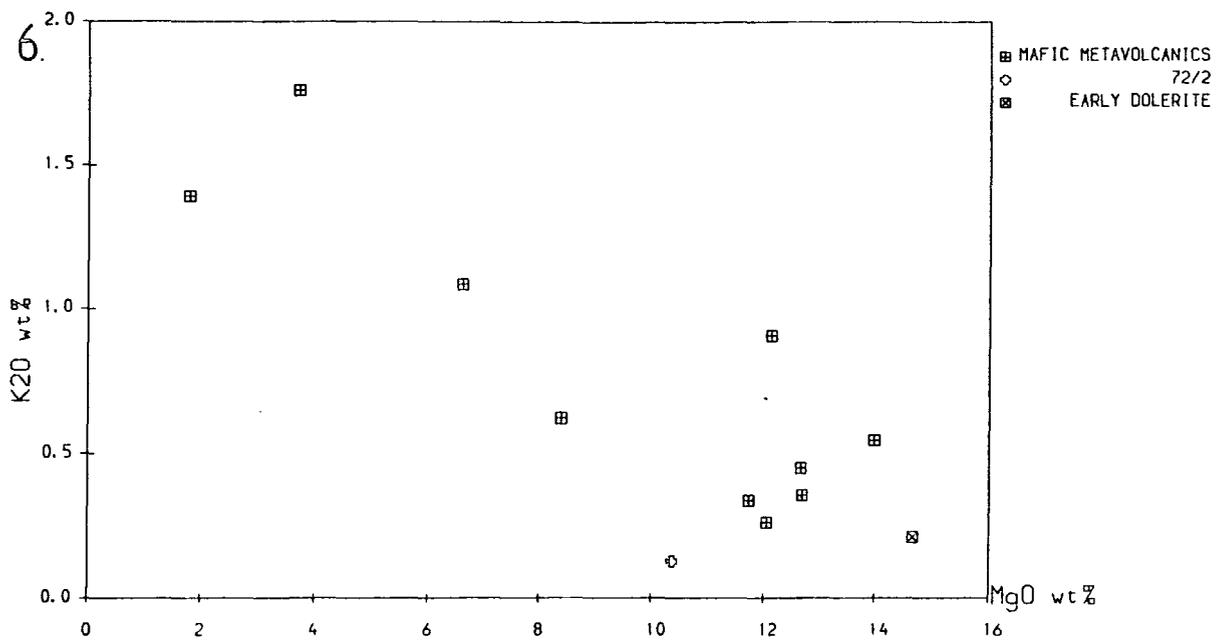
Mg contents ( and by implication, for the degree of fractionation which they have undergone ). In Ardnamurchan these high - Ti rocks range from picritic basalts with as much as 14.7% MgO to evolved ferrobasalts and trachyandesites; the Cuingleum intrusion described below ( section 2.4.3 ), which includes granitic rocks, may also form part of the plateau basalt suite. The suite is mainly represented by amygdaloidal basalt to picritic basalt lava flows. In the exposures examined, all in western Ardnamurchan adjacent to the Hypersthene Gabbro, the lavas are mainly composed of picritic basalt: out of 11 analysed basic rocks from this suite, 7 have more than 11.5% MgO. With one exception, there is no petrographic evidence for accumulation of olivine phenocrysts in these rocks, although several of the rocks have been heavily altered and subsequently hornfelsed and any olivine phenocrysts originally present would be difficult to identify as such. Sample 72/2 ( see below and Appendix 2 ) is particularly badly altered; 48/1, the most magnesian rock of all, is intensely hornfelsed but appears to have been relatively fresh prior to thermal metamorphism ( see below ). Sample 251 contains granular pyroxene and olivine aggregates which may be pseudomorphs after rare but large olivine crystals: it is also atypical in having a high Fe content ( see Fig. 2.5 ). Olivine is in general rather rare in these rocks, the dominant ferromagnesian mineral being a titaniferous magnesian augite. Apart from possible feldspar porphyrocrysts in 48/1 and adjacent basic hornfelses and in the ferrobasaltic sills, the rocks appear to be aphyric. The entire sequence is at most a few hundred metres thick at the present time, although this may be more a result of subsequent uplift and erosion than an indication of the original thickness of the lava pile. At present, the lavas are only exposed in western Ardnamurchan in widely separated downfaulted blocks ( see Map 1 ) and represent the dissected remnants of what was originally a more extensive lava field. In addition to the extruded rocks, a number of early Ti - augite bearing minor intrusions cut the Jurassic sequence, notably in the floor of the Dubh Chreag Gorge ( ferrobasaltic sills: Grid ref. 454631 ) and on the coast west of Sron Bheag in the form of a number of dykes which run parallel to the regional dyke swarm trend in this area ( strike 140 - 150° ). In addition, many of the picritic rocks within the lava sequence are medium - grained, ophitic rocks and may be sills intruded into the lava pile or the massive interiors of thick flows: it was not in general possible to identify flow margins within the small outcrops available. Less regular intrusions belonging to the plateau basalt suite such as 48/1 ( see below ) are present in the Glebe Hill area ( see Fig. 3.19 ). These appear to be inclined sheet intrusions, with northward dips, and may indicate the involvement of plateau basalt type magmas in the earliest stages of the development of the central complex in Ardnamurchan, as is the case in Skye ( Dickin et al. 1984 ).

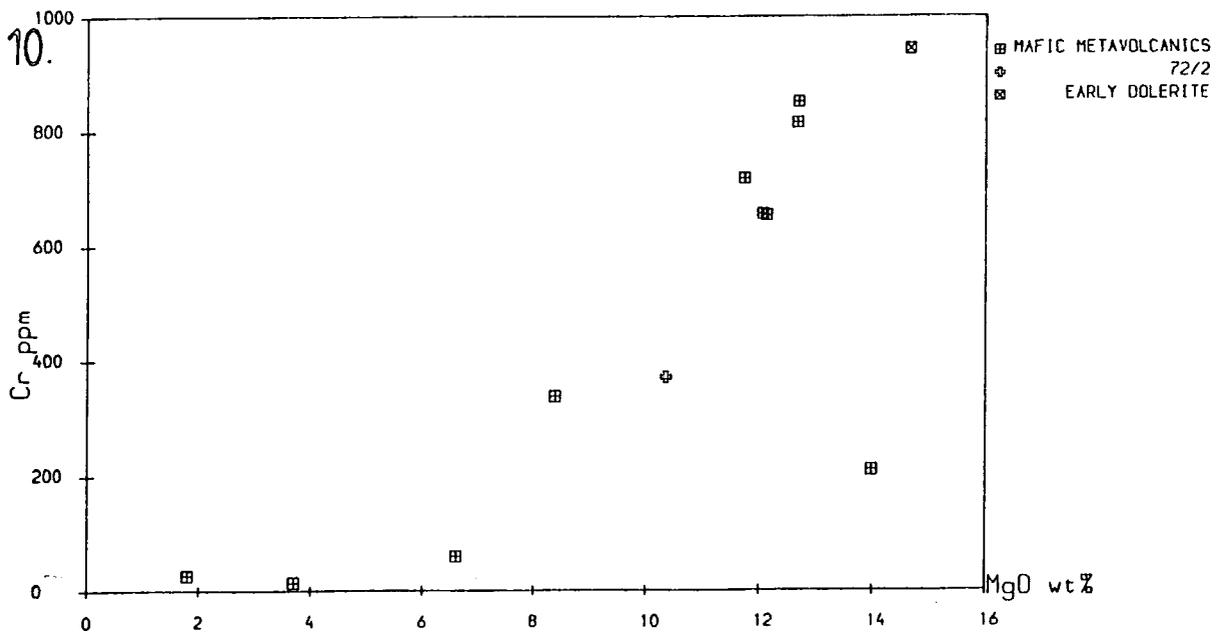
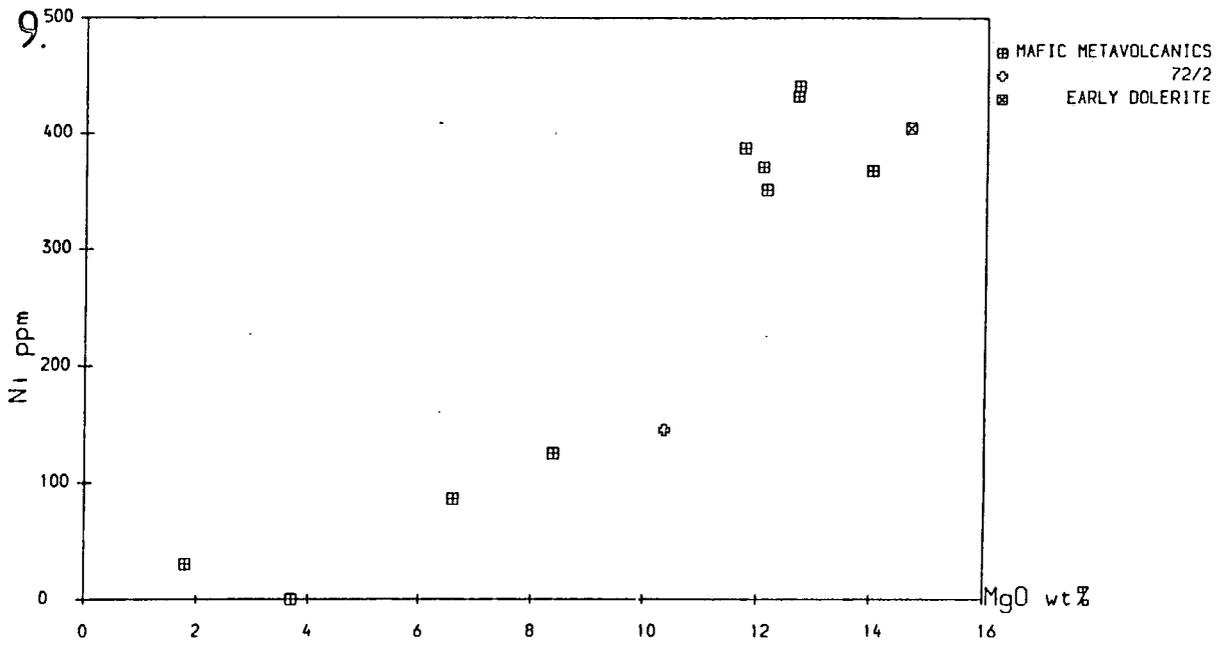
Fig. 2.6. Geochemical data for the early Plateau Basalt suite in Ardnamurchan.

72/2 is a plagioclase - olivine - orthopyroxene hornfels from within the plateau basalt hornfels at the northern end of Glebe Hill ( see Map 1 and section 3.2.6 ) which appears to have been an intensely weathered basic rock prior to thermal metamorphism.









Where the base of the lavas is exposed they overlie a thin sequence, no more than a few metres thick, of Tertiary sediments. Apart from a thin sandstone unit in the east of the central complex, these consist of organic - rich mudstones, lignite and altered basaltic tuff ( Richey et al. 1930 ). These rocks are exposed in the eastern part of the complex and at the base of the basalts at Sron Bheag ( see Map 1 and Fig. 3.17 ) and may also correspond to the banded hercynite - plagioclase - corundum - ilmenite hornfelses exposed adjacent to the contact at the north end of Glebe Hill. The presence of abundant organic material in the original sediment would certainly explain the remarkably reduced condition of these hornfelses ( see Chapter 5 ). During the course of the present work these hornfelses were found to form a series of south - dipping tabular bodies, broadly conformable to the adjacent hornfelled basalts. This conclusion is similar to the interpretation of the field geologists of the British Geological Survey ( Richey et al. 1930, p.104 ), who considered that the spinel hornfelses formed a screen between basic hornfelses and the Hypersthene Gabbro, and contradicts that of H.H.Thomas in the same work ( p.235 ), who considered that the spinel hornfelses were xenoliths in the Hypersthene Gabbro.

The interpretation of the geochemistry of these early basaltic rocks is complicated by the effects of alteration. However, for the principal purpose for which they were analysed during the present work, that of investigating possible contaminants of Hypersthene Gabbro magmas, the effects of the alteration can in a sense be ignored because the main, pervasive hydrothermal metamorphism to affect these rocks occurred prior to intrusion of the Hypersthene Gabbro ( section 3.2.5.5 ). Thus their present - day composition is probably a close approximation to their composition at the time at which they could have been assimilated, given a hot enough magma to melt them or sufficient time for diffusional equilibration to occur. The majority of the plateau basalt suite rocks analysed are too magnesian and hence too refractory to be likely contaminants but the ferrobaltic sills in particular are potential contaminants, having a liquidus temperature of about  $1100^{\circ}C$ , according to the liquidus T - MgO relationship of Thompson ( 1973 ) ( see Chapter 4 for discussion of the intrusion temperatures of the Hypersthene Gabbro magmas which were in contact with these rocks ).

Various features of the geochemistry of the plateau basalt suite in Ardnamurchan are summarised in Fig. 2.6. The one low - Ti basalt, which is also low in Zr and other incompatible elements, is also plotted on this diagram. This rock ( 72/2 ) is intensely hornfelled and contains granular

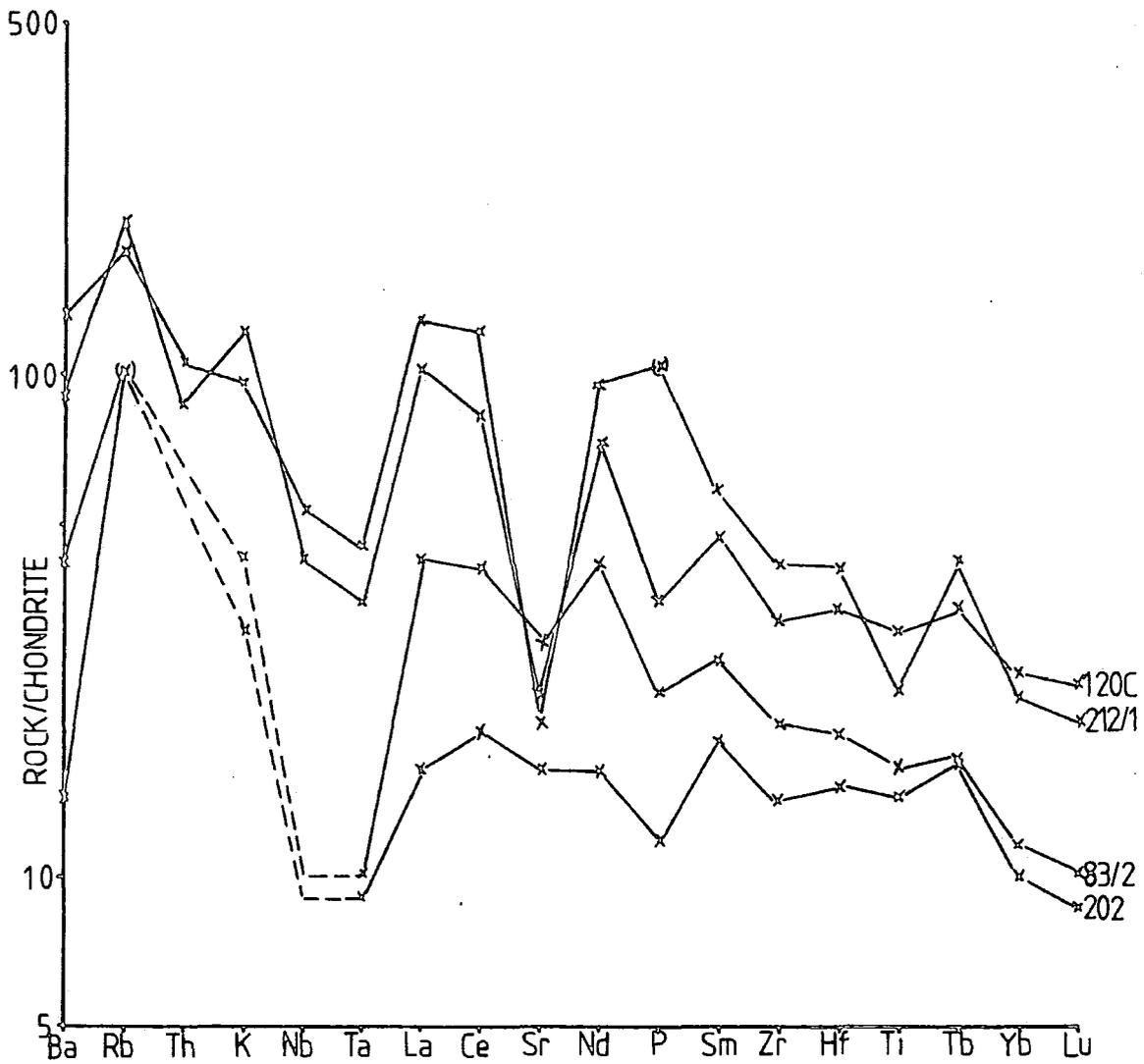
orthopyroxene pseudomorphs after mafic phases, suggesting that it has been intensely altered: however, the low abundances of immobile elements ( e.g. Ti and Zr ) in this rock suggest that it may have originally been an incompatible - element poor tholeiite. Apart from this one sample, the suite is geochemically comparable to the corresponding plateau basalt suites in Skye and Mull, the Skye Main Lava Series and the Mull Plateau Group, although characterised by more magnesian, primitive compositions. Much of the scatter in the plots shown in Fig. 2.6 can be attributed to the following causes:

1). Alteration. Studies of the geochemical effects of low - temperature sub - aerial or freshwater alteration of basalts in the British Tertiary Province ( Morrison et al. 1980 ) and in Iceland ( Wood et al. 1976 ) indicate that Group 1 and 2 elements will be most strongly affected by such alteration: this is consistent with the considerable scatter in K, Na, Ca and Sr apparent in Fig. 2.6.

2). Crustal contamination. This is most apparent in the plot of Zr against MgO, where although the low - MgO samples lie on a linear fractionation trend the picritic basalts tend to lie above the trend, with higher Zr contents. The studies above concluded that Mg was relatively immobile except under extreme alteration whilst Zr is generally considered to be completely immobile. The most plausible explanation of this plot is that the high - Mg rocks contain a greater proportion of Zr - rich crustal contaminant than the parental magma to the fractionated basalts. The occurrence of crustal contamination in these rocks is also apparent from Fig. 2.7, a chondrite - normalised trace element plot showing two of the evolved basalts, a basaltic trachyandesite, and a picritic basalt. All four show depletion in Nb and Ta relative to their abundances of K and La. This is a characteristic feature of those British Tertiary basic rocks that have been contaminated with the calc - alkaline rocks of the Lewisian or with supracrustal sediments. Within the uncertainty of the data, Hf/Sm and Zr/Sm are normal, excluding major contamination by sedimentary rocks of the Moine or Mesozoic, or by the most extremely fractionated Lewisian rocks. A slight Th depletion may be present in some of the rocks ( Rb is likely to have been affected by alteration - see above - and cannot be used to indicate anything about previous contamination ), suggesting that although the dominant contaminant is likely to have been amphibolite - facies Lewisian material, some granulite - facies rock has also been assimilated. This point is discussed further in section 4.3.2.

The remaining plots in Fig. 2.6 show evidence for the operation of fractional crystallisation. The fall

2.7 Chondrite - normalised minor and trace element data for Plateau Basalt Suite rocks: picritic basalt ( 202 ), basalt ( 83/2 ), ferrodolerite sill ( 120C ) and basaltic trachyandesite ( 212/1 ).



in  $\text{TiO}_2$  with decreasing MgO below about 4% MgO suggests entry of titanomagnetite at that point, whilst inflections in the  $\text{TiO}_2$  and  $\text{Al}_2\text{O}_3$  plots at about 10% MgO mark the entry of plagioclase into the fractionating crystal assemblage. Prior to this, the rapid fall in Cr relative to Ni, the relatively small variation in Ni with decreasing MgO and the relatively constant value of  $\text{TiO}_2$  in the picritic basalts suggests that the early fractionation of these rocks may have been dominated by titaniferous augite and possibly orthopyroxene, rather than olivine ( which would remove Ni relative to Mg and Cr ). This implies that the variation in the picritic basalts must have been produced by high - pressure fractionation, at pressures of 10kb or so where cpx  $\pm$  opx are stable liquidus or near - liquidus phases. Unfortunately, CaO does not show a coherent trend with MgO at high MgO, presumably because of alteration, so clear evidence for or against early clinopyroxene fractionation in these rocks is lacking. Conversely, the inflections in the plots of  $\text{Al}_2\text{O}_3$  vs. MgO and  $\text{TiO}_2$  vs. MgO at about 10% MgO imply the early appearance of plagioclase in the fractionating crystal assemblage which produced the more fractionated rocks and, therefore, production of the more fractionated plateau basalt suite rocks by low - pressure crystal fractionation.

#### 2.4.2. Volcaniclastic Rocks.

Unlike the other two igneous centres defined by Richey et al. ( 1930 ), Centre 1 of the Ardnamurchan central complex is dominated by volcaniclastic rather than by plutonic rocks. These do not occur at the contact of the Hypersthene Gabbro at the present day but may have formed much of its eastern margin in particular, prior to the emplacement of Centre 3 of the complex. These rocks were interpreted by Richey et al. as intrusive, vent - fill agglomerates, being mainly composed of coarse, chaotic breccias which in some cases had steep contacts against adjacent rocks. However, a number of field observations made at the nearest outcrops to the Hypersthene Gabbro, in a north - south cliff which forms the eastern side of Glendrian Bay ( Grid Ref. 461705/461708 ) suggest that these outcrops at any rate are in fact extrusive in origin:

- 1). At the southern end of the cliff, near its base and accessible from below, shallow channels filled with bedded epiclastic pebbly sandstone occur in coarse volcaniclastic breccias.
- 2). These breccias show a crude horizontal stratification further along the cliff, defined by coarser and finer - grained units several metres thick. There is no evidence of steep, intrusive or cross -

cutting contacts between breccia units.

3). The breccias are composed of, variously, clast- and matrix-supported rocks with a very fine-grained, often silty or muddy matrix ( now pervasively hornfelsed ) and a completely unsorted clast assemblage, with clasts up to several metres across. There is no evidence for elutriation of fine-grained material such as normally occurs in volcanic vents.

4). The clast assemblage is polymict, with a great variety of felsic to basic, mainly fine-grained, igneous rocks, calcareous sandstone blocks and shaly debris, but no recognisable Moinian sediments, such as would be expected in vent structures since Moinian rocks probably occur only a few hundred metres below these outcrops ( section 3.2.2, above ). There is no evidence for prior hornfelsing of any of these rocks, or for juvenile chilled igneous fragments.

It therefore seems likely that these are reworked epiclastic deposits. The chaotic, poorly sorted and mud-rich character of the rocks suggests that they are some type of laharic or mudflow deposit, with localised reworking by surface streams to produce the bedded sandy deposits. The presence of abundant Jurassic clasts suggests that active erosion of exposed Jurassic rocks within the volcanic edifice was taking place.

Texturally similar but compositionally rather different rocks were also examined briefly at Faskadale, about 4km further east, during the course of sampling in that area, and the same conclusions regarding their mode of origin apply. If the bulk of the 'vent agglomerates' are in fact extrusive in origin, as has been suggested by other workers ( R.N.Thompson, pers comm. ), then, although these rocks may have formed much of the walls and roof of the Hypersthene Gabbro above the present level of exposure, they cannot form the wall below the present-day surface except perhaps on the missing eastern margin of the pluton. There is some indirect evidence to suggest that volcanoclastic rocks may have formed the wall rocks of the intrusion at the present level of exposure on this margin. The present-day outcrop of the 'Centre One' volcanoclastic rocks is roughly elliptical in plan and must be steep-sided, because it cuts across as much as 200m of topography. Richey et al. ( 1930 ) describe a number of steep contacts between volcanoclastic rocks and Pre-Tertiary rocks on the eastern side of Centre 1, as well as flat-lying contacts. For the most part, however, the present-day northern and southern margins of Centre 1 are formed by arcuate intrusions of felsite,

microgranite and microgabbro ( see Fig. 1.2 and section 2.4.3, below ). Volcaniclastic rocks of the same age do however occur outside this area, notably at Glendrian Bay ( Grid Sq. 4670 ) in a block downfaulted between the Glas Eilean fault and a fault concentric to the Hypersthene Gabbro ( see section 3.2.2 ), suggesting that the original extent of the volcaniclastic rocks was much greater. One possible interpretation of these observations is that Centre 1 is a caldera structure, in which case the eastern side of the Hypersthene Gabbro would be inside the earlier Centre 1 caldera and against volcaniclastic rocks which had previously been substantially downfaulted relative to those around the rest of the pluton.

Only two samples of these rocks were collected for geochemical analysis. These approximate to felsic ( 190 ) and intermediate to basic ( 301/1 ) igneous rocks respectively ( see analyses in Appendix 2 ). 190 in particular does however show the effects of surficial weathering, particularly in its anomalously low Na/K ratio and Na content: this does not seem to reflect addition of K ( see Fig. 2.8, below, where K lies on a smooth curve through the chondrite - normalised abundances of the other incompatible elements ) but leaching of sodium instead.

#### **2.4.3. Early Tertiary basic to felsic intrusions.**

In addition to the small sills and dykes associated with the basaltic lava pile, a number of slightly larger early intrusions occur close to the margins of the Hypersthene Gabbro. These are mainly associated with Centre 1 of the central intrusive complex, forming a series of arcuate intrusions around its periphery and shallowly dipping sheet intrusions ( cone sheets and the Beinn an Leathaid hybrid granophyre - dolerite sheet ) within it ( see Fig. 1.2 ). In addition, a number of mixed acid - basic intrusions occur on the southern margin of the pluton, one of which is a vent structure with a strong NNW - SSE elongation, parallel to the regional dyke swarm. The felsic members of this early intrusive suite are particularly significant because they have near - eutectic compositions and hence are particularly prone to high degrees of partial melting and remobilisation as magmas when reheated by later intrusions. Such recycling of felsic intrusive rocks within the British Tertiary central complexes has been postulated by Thompson ( 1982 ) on the basis of the irregular pattern of fractionation with time observed in the granitic complexes, particularly on Skye. Similar remobilisation of earlier felsic rocks has been identified in certain recent central volcanic complexes ( notably Askja, Iceland ( MacDonald et al. 1987 ) and Yellowstone in the northwestern United

States ( Taylor 1988 ) on the basis of chemical and isotopic depletions in fresh quenched magmas attributable to hydrothermal alteration prior to remelting and incorporation of pre - existing felsic rocks into these magmas.

Pre - Hypersthene Gabbro Tertiary intrusive rocks were sampled in two areas. The first of east of the northern side of the Hypersthene Gabbro, in the Faskadale area, where felsite and microgranite intrusions occur in the arcuate belt of intrusions which separate the volcanoclastic rocks of Centre 1 from the Jurassic rocks which form the northern coast of Ardnamurchan in this area ( see Fig. 1.2 ). These include a felsite ( 299 ) with sparsely distributed mafic blebs and showing intense hydrothermal alteration with the development of sericite, chlorite, carbonate and pyrite, and a microgranite ( 300/1 ) containing diffuse - edged rounded quartzite blocks and showing hydrothermal alteration, with the development of the same assemblage as is present in 299.

The second group of early felsic rocks sampled come from the south coast of Ardnamurchan in the Maol Bhuidhe area. Early felsic intrusions in this area, whose age relationships to the Hypersthene Gabbro are indicated by the presence of pre - Hypersthene Gabbro alteration in the rocks concerned ( corresponding to the M0 phase of metamorphism ( section 3.2.5.5 ) ), consist of two or more mixed acid - basic sheets, probably irregular sills, around 470627 - 468626 and a variety of felsic rocks in the elongate vent - intrusion at Cuingleum ( Grid ref. 464623 ). The latter is an elongate NNW - SSE orientated plug with a microgabbroic core and marginal felsitic and microgranitic intrusions emplaced into Tertiary basalts and middle Jurassic sediments downfaulted into a vent - like structure along inward - dipping faults. The felsites are cut by tuffitic pipes and dykes, filled with rounded clasts of felsite, and with brecciated margins, implying that the intrusion was the site of explosive activity. The felsic rocks are extensively remelted at their contact with the gabbro plug, producing a suite of hybrid rocks. This implies that the gabbro plug, which is at most only 50m wide, was a magma conduit through which large amounts of basic magma flowed: since it lies on the regional dyke swarm trend and is not radial to Centre 2 of the central complex, it may in fact be a feeder to the upper part of the plateau basalts in the area: there is however no geochemical data for the microgabbros to prove or disprove this. The analysed early felsic rocks from the margin of the vent ( 212/3 ) and the sheet intrusion to the east ( 207 ) have slightly higher Zr ( 411 and 452 ppm respectively ) than felsic rocks with similar degrees of fractionation ( as indicated by Ba, Sr and SiO<sub>2</sub> contents ) from the central complex ( see Chapter 4 ), suggesting a more alkaline character prior

to alteration, because of the increased solubility of Zr in alkaline melts ( E.B.Watson & Harrison 1983 ). This would be consistent with production of these rocks by fractionation ( with or without assimilation ) from the more alkaline magmas of the plateau basalt suite rather than the tholeiites of the central complex ( see Chapters 4 and 6 ).

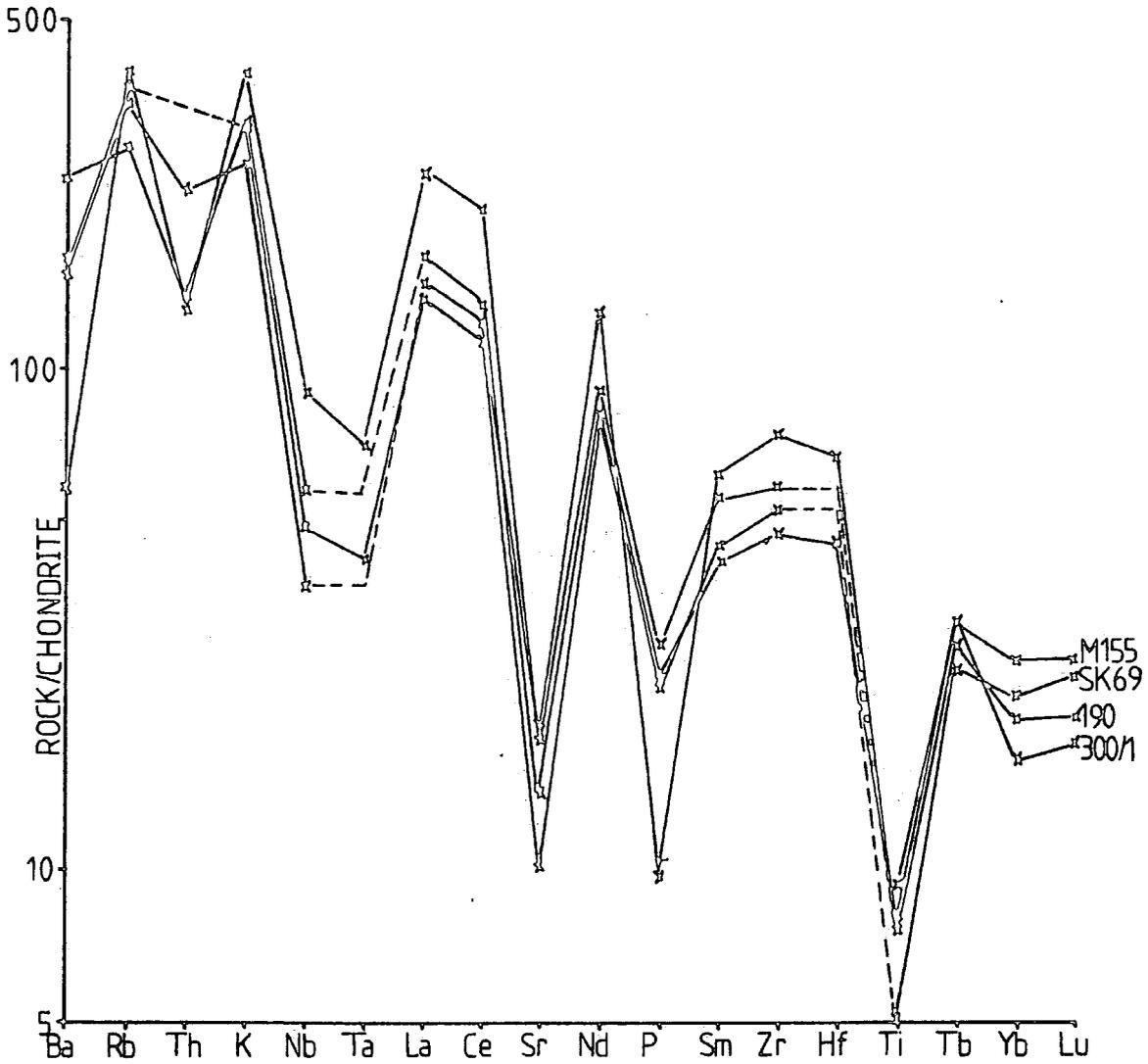
The trace element geochemistry of two Centre 1 felsic rocks, the epiclastic sandstone 190 and the xenolithic microgranite from Faskadale ( 300/1 ), is compared in Fig. 2.8 with that of two typical relatively unfractionated granites from Skye and Mull. The four rocks are basically very similar, suggesting that the origin proposed for the Skye and Mull granites by Thompson ( 1982 ), by fractionation of a tholeiitic magma contaminated with felsic Lewisian gneisses, also holds for the early, pre - Hypersthene Gabbro rocks in Ardnamurchan as well. The contamination history of these rocks is discussed further in section 4.3.2, along with that of the rocks from the outer part of the Hypersthene Gabbro. It is however apparent from the slightly elevated Hf/Sm ratio of 300/1 in particular ( see Appendix 2 and Fig. 4.21 ) that it has been contaminated with Lewisian granitic rocks or with Moinian or Mesozoic metasediments. The presence of corroded quartzitic xenoliths in 300/1 makes the latter alternative more likely.

Although there is a significant amount of geophysical data ( Bott & Tuson 1973; Barrett 1987; Harrison 1987 ) relating to the present - day subsurface extent of the Ardnamurchan central complex as a whole ( see Chapter 7 ) it is difficult to use this as anything other than an upper limit on the size of near - surface Pre - Hypersthene Gabbro intrusions. The much larger volume of exposed later intrusions, relative to that of the earlier intrusions, suggests that only a small proportion of the mafic plutonic rock detected geophysically beneath Ardnamurchan predates the Hypersthene Gabbro. However, it is not valid to extrapolate from the relative volumes of exposed rocks of different Centres to those of buried plutonic rocks in Ardnamurchan, because it is apparent from the presence of volcanoclastic rocks in Centre 1 that it is much less deeply eroded.

## **2.5 The Pre - Hypersthene Gabbro Structure of Western Ardnamurchan.**

Most of the large - scale structures present in western Ardnamurchan have been discussed in preceding sections of this chapter. These are summarised here by age, together with a discussion of smaller - scale structures and fabrics in the various groups of pre - Hypersthene gabbro rocks which

2.8 Chondrite - normalised minor and trace element data for a Centre 1 microgranite ( 300/1 ) and a Centre 1 agglomerate, compared with the Glamaig granite, Skye ( SK69 ), and the Derrynaculen Granophyre, Mull ( M155 ).



SK69 from Thompson 1969, Thorpe et al. 1977 and Tammenagi 1976

M155 from Walsh et al. 1979

may have affected their modes of deformation.

Much of the crust beneath Ardnamurchan is probably formed by plutonic and high - grade metaplutonic rocks, of the Lewisian gneiss complex for the most part but also possibly including Grenvillian rocks and Tertiary plateau basalt suite magma chambers in the very deepest part of the crust. At the surface, in Coll and Tiree, the Lewisian rocks have a steep, roughly north - south trending fabric defined by foliation and gneissic banding ( Richey et al. 1930 ). A number of steeply inclined, mainly NNE - SSW trending brittle faults and greenschist facies fault zones were also described from the islands by these workers. Overall, however, the deep seismic reflection data of the BIRPS group ( Smythe et al. 1982; Blundell et al. 1985 ) indicate that there are few subhorizontal reflectors in much of the Lewisian basement in north - west Scotland, and by inference few structurally contrasted layers which might promote flexural deformation or act as stress guides ( Gudmundsson 1988 ).

In contrast, the deformed rocks of the Moine Thrust Zone, where exposed to the north of Ardnamurchan, and the rocks of the Moine Group itself, are characterised by strong structural anisotropy. West of the steep folded belt at the eastern edge of the Tertiary complex in Ardnamurchan, the Moinian rocks have only shallow, mainly eastward dips: the same is also true of the Moine Thrust Zone at its western edge ( Coward 1983 ; Soper & Barber 1982). The rocks are micaceous and have a well - defined schistosity in this section, particularly in the case of the semipelites. This foliation and the alternation of psammitic and semipelitic units means that these rocks are likely to deform by bending and folding, since they form a multilayered structure with weak layers ( the semipelites ) alternating with stronger layers. The NNW - SSE trending isoclinal fold belt to the east of the Tertiary complex is also of note as the earliest example of structures with this orientation in the area ( although this does not necessarily imply that later structures simply follow Moinian planes of weakness: see Chapter 7 ).

The Mesozoic rocks of Ardnamurchan do not seem to have been greatly deformed prior to the Tertiary, although they are cut by NNW - SSE trending ( strike  $160^{\circ}$  approx. ) faults, one of which, the Glas Eilean fault, seems to have been active since the early Jurassic at least, as a westward - downthrowing fault defining the edge of the Inner Hebrides sedimentary basin in Ardnamurchan. The rocks themselves are well - bedded and laminated, and the pelitic units commonly have a shaly fissility. Pre - Hypersthene Gabbro Tertiary metamorphism of these rocks accentuated this fissility

through the development of a pervasive micaceous schistosity ( see section 3.2.5.5 ). These rocks would therefore be expected to show a strong tendency to develop flexural slip and hence deform by bending and folding, at least prior to the high - grade contact metamorphism which affected many of them immediately after the initial emplacement of the Hypersthene Gabbro ( section 3.2.5.5 ).

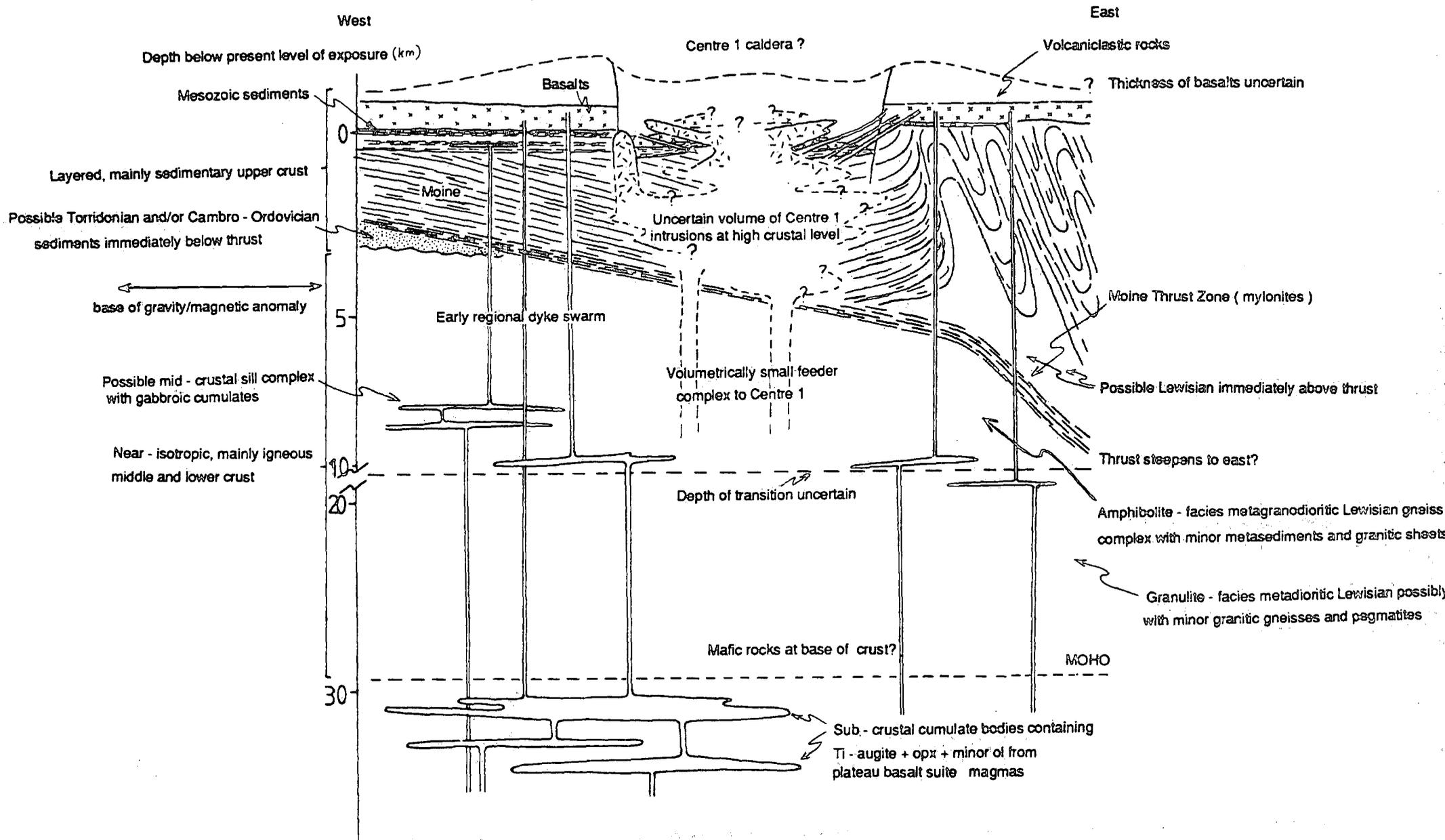
The Tertiary plateau basalt sequence in Ardnamurchan is also downthrown to the west by movement on the Glas Eilean fault, as are Centre 1 volcanoclastic rocks near Glendrian Bay ( Grid Square 4670 ). It is not clear whether this movement occurred prior to, or after, initial emplacement of the Hypersthene Gabbro: there is clear evidence for syn- or post- Hypersthene Gabbro movements on this fault but movements could also have occurred earlier in the Tertiary as features of the Mesozoic succession indicate that the fault was in existence prior to the Tertiary. The main pre - Hypersthene Gabbro structures of Tertiary age are the NNW - SSE trending ( strike  $140 - 150^{\circ}$  ) dyke and elongate plug swarm which also outcrops in Mull to the south and Muck to the north - west, and the possible Centre 1 caldera structure. The dyke swarm is one of a number of such swarms which occur up the west coast of Scotland and extend as far south and east as northeast England. They indicate a dominantly extensional regional stress field with WSW - ENE extension, although the presence of north - south trending subsidiary dyke swarms and detailed studies of dyke dilations suggest that a component of dextral shear may also have operated during at least part of the period of igneous activity in the British Tertiary Province ( England 1988 ). In Ardnamurchan, these dykes are associated with a number of faults, also with NNW - SSE orientations: these are discussed further in Chapter 3. The basalts and picritic basalts themselves, at the present level of exposure, form an apparently continuous sequence of massive to amygdaloidal, variously altered to hornfelsed lava flows. However, weathering of flow tops, a common phenomenon in the British Tertiary province as a whole, is generally obscured by metamorphism close to the central complexes and may have given these rocks a mechanically layered structure prior to, and in the initial stages of, the emplacement of the Ardnamurchan central complex. In addition, the occurrence of felsic rocks and explosive brecciation in the vent intrusion at Cuingleum suggests that the basaltic rocks above the present level of exposure may have been overlain by, or interbedded with, pyroclastic rocks. This would also produce mechanical layering in the Tertiary extrusive sequence.

The agglomerates of Centre 1 are only slightly layered or stratified, but the presence of cone sheets and larger subhorizontal intrusions within them suggests that the partly extrusive, partly intrusive

sequence of Centre 1 will behave as a layered unit overall. The greatest uncertainty regarding Centre 1 is, however, the abundance of high - level intrusive rocks beneath it and adjacent areas, although the geophysical evidence does indicate that these are unlikely to extend to a depth of more than 3 - 5km below the surface, apart from volumetrically minor magma feeder structures ( see Chapter 7 ).

A possible overall model for the composition and structure of the crust beneath Ardnamurchan immediately prior to the emplacement of the Hypersthene Gabbro is presented in Fig. 2.9. There are many uncertainties associated with this, of which the most critical to the present work are the precise position of the Moine Thrust Zone ( the boundary between the meta-igneous middle and lower crust and the layered metasedimentary upper crust ) and the vertical and horizontal extent of pre - Hypersthene Gabbro Tertiary intrusive rocks. The position and geometry of the plateau basalt suite intrusions in the middle crust is speculative and based on analogy with geophysically - detected intrusions in analogous magmatic provinces such as the Socorro section of the Rio Grande rift in south - western North America ( Rinehart et al. 1979; Iyer 1984 ). The low - pressure fractionation trend shown by the more evolved members of the plateau basalt suite in Ardnamurchan does however require the presence of such intrusions in the upper half of the crust.

## 2.9 Schematic section through the Crust and Uppermost Mantle beneath Ardnamurchan prior to emplacement of the Hypersthene Gabbro. See text for data sources



### 3. FIELD AND PETROGRAPHIC EVIDENCE FOR A REVISED SUBDIVISION OF THE HYPERSTHENE GABBRO AND FOR THE AGE RELATIONSHIPS OF ROCKS, METAMORPHISM AND STRUCTURES IN AND AROUND THE PLUTON.

#### 3.1 An Overview.

As was discussed in Chapter 1, the Hypersthene Gabbro has always been considered to form a single pluton, albeit with numerous associated minor intrusions. This is because of its characteristic age relationships with the other plutons which were defined by Richey et al. ( 1930 ), the lack of well - developed chilled margins within it which would unambiguously define separate intrusions, and the distinctive orthopyroxene - rich mineralogies of most of the rocks of which it is composed. More recent models of its geometry and emplacement history ( Wells 1978, Walker 1975 ) advocate the presence of multiple intrusions within the Hypersthene Gabbro but retain the assumptions that a single emplacement mechanism can account for all parts of it and that the multiple intrusions which make up the pluton are all of essentially the same type.

This chapter presents a revised sub - division of the Hypersthene Gabbro, together with evidence for the geometries of its various components ( detailed discussion of the emplacement mechanisms involved is deferred to Chapter 7 ). Aspects of the petrography and mineralogy of the igneous rocks in and around the pluton are described which, together with cross - cutting relationships seen at outcrop and the age relationships of metamorphism and deformation in the country rocks, are used to construct a detailed relative chronology of events and extended periods of activity in the evolution of the pluton as an active subvolcanic magma reservoir. This chronology will be used in subsequent chapters as the framework for discussion of the rocks of the Hypersthene Gabbro and its contact aureole, and of the processes involved in their formation.

In this chapter the intrusion is divided into two main parts. The older of the two is the Marginal Border Group ( section 3.2 ) which forms most of the outer margin of the intrusion ( see Map 1 ) and has a complex contact zone adjacent to the country rocks and the minor intrusions ( mostly of cone sheet type ) associated with the Hypersthene Gabbro: relevant age relationships in these rocks are also discussed in section 3.2. The Marginal Border Group corresponds in part to the marginal

quartz dolerite facies identified by Richey et al. ( 1930 ) but also includes quartz - free and olivine - bearing microgabbros found further within the pluton. The second main component of the pluton is the Inner Series, whose contacts ( partly intrusive and partly structural or tectonic ) with the Marginal Border Group are discussed in section 3.3, as are criteria for distinguishing rocks belonging to each part of the pluton. The internal organisation of the Inner Series is discussed in section 3.4.

## 3.2 The Marginal Border Group ( MBG ).

### 3.2.1 Outcrop distribution and characteristics.

The outer contact zone of the Marginal Border Group tends to be rather poorly exposed, particularly inland, because of hydrothermal alteration of the rocks which has reduced their resistance to erosion. However, extensive coastal exposures of the contact zone occur on the wave - cut platform found around high tide mark along the northern coast of Ardnamurchan between Glendrian Bay and Sanna Point ( Fig. 3.1 and Maps 1,2 and 3 ) and these water - washed ( albeit barnacle - encrusted! ) outcrops provide abundant information on age relationships and meso - scale igneous structures in the contact zone. Elsewhere, the inner parts of the MBG are well exposed in the cliffs and crags between Ardnamurchan Point and Gharblach Mhor, although the contact itself lies offshore. Except around the inlets of An Acairseid, the sectors of the contact running from Druim na Cloise through Beinn nan Codhan and Tom na Moine to Kilchoan are sparsely and poorly exposed, but the overall geometry of the contact can be deduced from the three - dimensional constraints provided by exposures in a series of large north - south gullies developed along faults, dykes and crush zones. Finally, exposure in the Glebe Hill area is relatively good: however, field and other evidence suggests that many of the rocks in this latter area which were previously considered to belong to the Hypersthene Gabbro ( Richey et al. 1930, Wells 1954 ) may in fact belong to later intrusions.

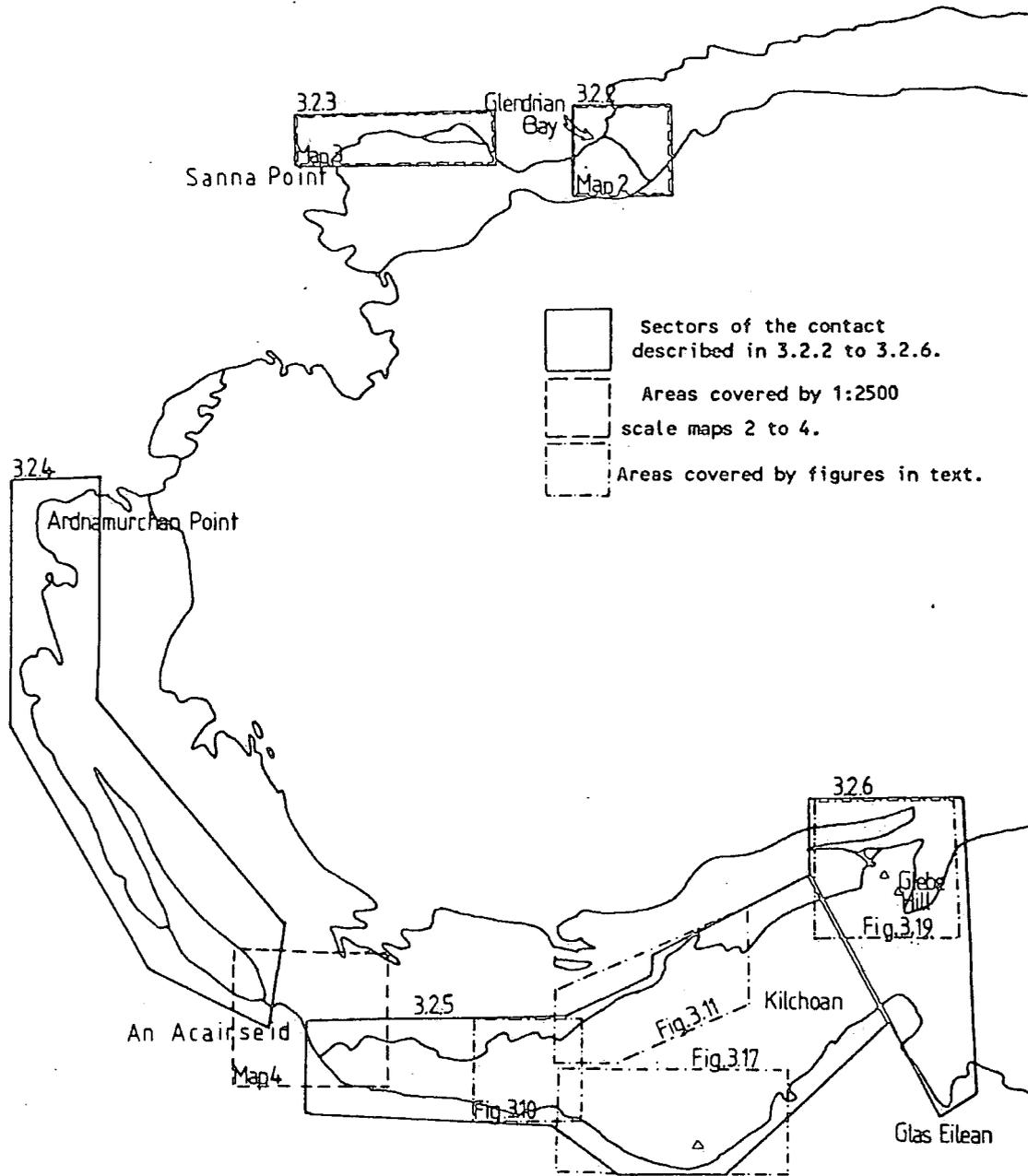
The various areas of interest are marked on Fig. 3.1. and will be discussed in anti - clockwise sequence, beginning with a detailed examination of the rocks on the northern side of the intrusion: the well - exposed geology of this area is frequently used in this chapter to interpret less well - exposed rocks elsewhere around the margin of the pluton.

### 3.2.2. Glendrian Bay and Hill 90.

#### 3.2.2.1. The outer contact of the Hypersthene Gabbro and the rocks of the Marginal Border Group.

Inland of Glendrian Bay, the contact runs N.N.W., from its truncation by the later Great Eucrite

Fig. 3.1 Location map for areas described in section 3.2.



to the south, along the western side of Hill 90. The contact itself is not exposed but outcrops of homogenous, rather granular microgabbro occur as little as 1 metre from its inferred position. These are cut by a few hornfelsed microgranitic veins which contain small, diffuse quartzitic inclusions. The latter suggest that the veins are backveins produced by melting of fusible lithologies in the country rock sediments. Alternatively, they may be related to the later net - veined intrusion further south on Hill 90, which appears to be an extension of the Sgurr nam Meann hybrid granophyre - quartz dolerite intrusion ( Day 1985; see Fig 1.2. ). However, the net - veined intrusion contains prominent feldspar phenocrysts and basaltic pillows and fragments, which are all absent from the microgranitic veins. In either case, there is no counterpart along this part of the contact of the complex contact zone found further north.

Within about 100m of the coast, a narrow zone of sparsely plagioclase - pyric microgranodiorite or granophyre is present between the gabbro and the country rocks. This widens to about 7m across on the wave - cut platform, around Grid Ref. 46007029. To the west, the contact bends sharply round to an east - west orientation and almost immediately passes out to sea. However, more plagioclase - pyric microgranodiorite is exposed near the low tide mark as far west as 45807030, in outcrops up to 20m wide ( see Map 2 and Fig. 3.3 ). The contact between the microgranodiorite and the rocks to the northeast is typically lobate or irregular but sharp, despite the fact that the latter are intrusive rheomorphic breccias ( see below ) all along the exposed part of the contact. The rheomorphic breccias do not contain microgranodiorite xenoliths. Although the microgranodiorite does contain a number of partially melted, felsitic - textured quartzofeldspathic xenoliths, these are not directly comparable with the rheomorphic breccias because they do not contain residual quartzite blocks ( see below ). There is, however, no evidence of chilling of either rock type at the contact between them. This suggests that, whichever of the two was intruded later, this occurred when the earlier intrusion was still hot and, therefore, that both were intruded prior to the abrupt cooling of the contact zone described below.

In thin section the microgranodiorites show a range of textures, but are characterised by bimodal ( or even trimodal, if the rare plagioclase phenocrysts are included ) grain size distributions. Typically, about 50% to 80% of the rock is composed of 1 to 2mm long plagioclase laths with a low length:width ratio, and smaller amounts of subophitic augite ( patchily altered to green amphibole and/or chlorite ) and equant opaque grains. The remainder of the rock varies from fine - scale granophyric quartz

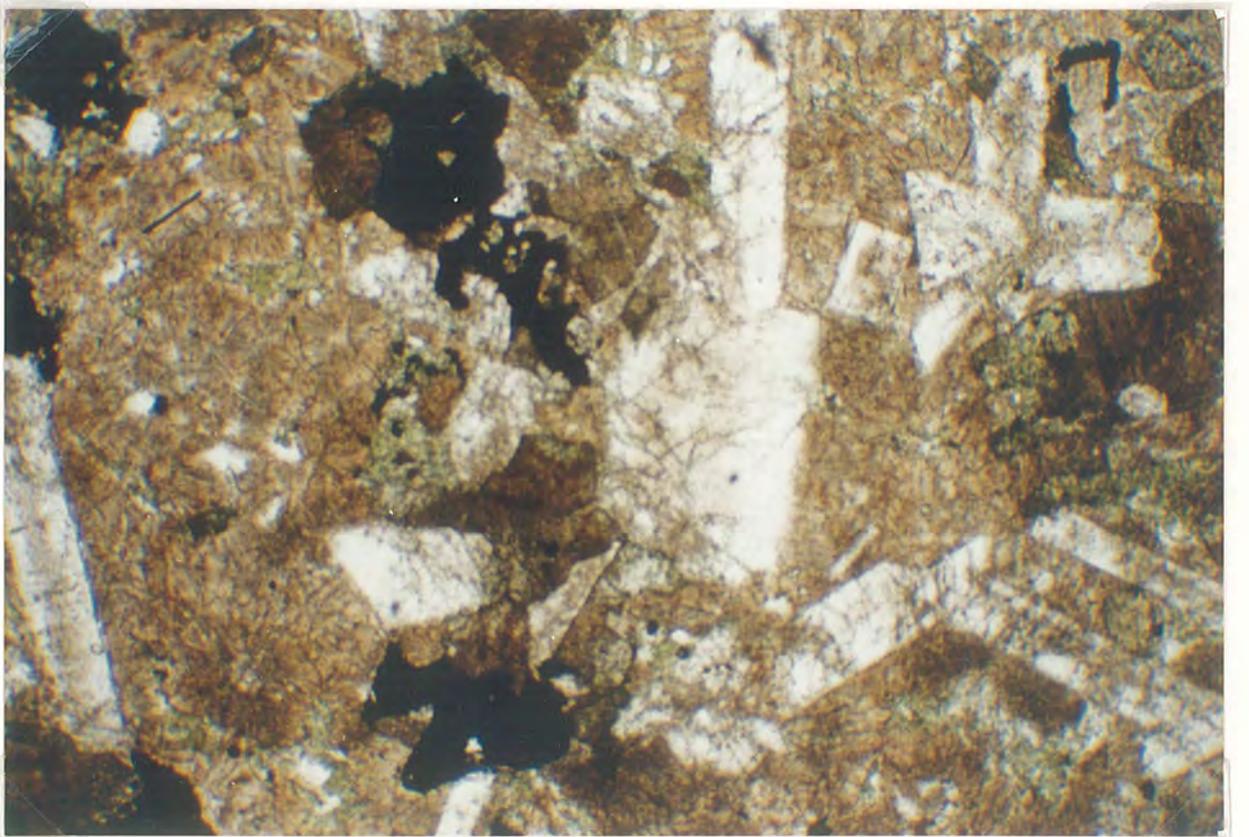
- feldspar overgrowths on plagioclase crystals, through variolitic material to spherulitic textured devitrified glass which occupies the interstices between the large crystals ( Plates 3.1, 3.2 ). The spherulites in particular indicate crystallisation under conditions of extreme supercooling, by as much as  $400^{\circ}\text{C}$  ( Lofgren 1971; see also Chapter 5 ). In some sections the spherulites show sharp spheroidal devitrification fronts ( Lofgren 1971 ) against aggregates of hydrothermal quartz, epidote and chlorite ( Plate 3.3 ). This suggests that devitrification of the quenched interstitial material was halted by the entry of cool hydrothermal fluids into the rocks, which resulted in the dissolution and replacement of the remaining metastable glass.

Although too few sections of these rocks are available to demonstrate or exclude any systematic variation in the devitrification textures with location, for example with distance from the contact, it seems that there is an overall pattern of relatively slow cooling, with crystallisation taking place under conditions of only very mild supercooling ( Lofgren 1974 ), being followed by much more rapid cooling which produced the very high degrees of supercooling needed to produce variolitic and spherulitic devitrification textures. The microgranodiorite is invariably isotropic, which rules out the possibility that it might have been emplaced as a crystal - rich mush, under which circumstances a flow fabric defined by alignment of feldspar laths would have developed ( Hutton 1988 ). This implies that the two - stage crystallisation history inferred from the texture of the granodiorite occurred in situ and reflects a drastic change in cooling rate, and therefore in rate of heat loss, during crystallisation.

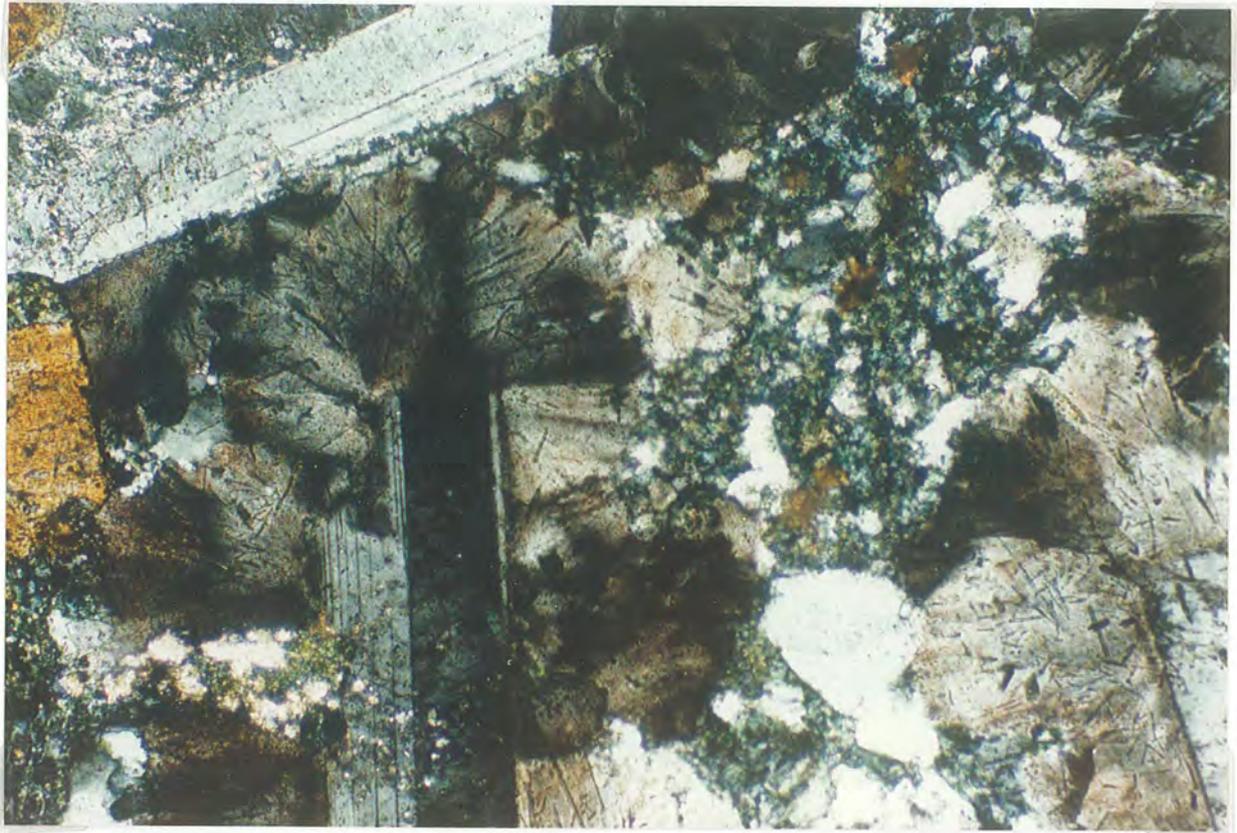
Between the Great Eucrite and the coast around Glendrian Bay, the contact is slightly uneven but is very steep or subvertical overall throughout a vertical distance of more than 50 metres ( Map 2 ). On the intertidal rock platform around the bay the maximum vertical exposure is no more than 4m. Within this distance the outer contact, between the microgranodiorite and the rheomorphic breccias, is steep, but the inner contact of the microgranodiorite in this area is much less regular. This latter contact is between microgranodiorite and a variety of basic and heterogenous hybrid rocks which are discussed below. Although steep in most places, around 45877030 the inner contact is subhorizontal, with microgranodiorites underlying microgabbros and hybrid rocks in an area up to 5m wide ( see Fig. 3.3 ). On a smaller scale, the contact is lobate or pillowed ( Plate 3.4 ). Lobes of fine - grained dolerite, with lobate or cauliform chilled basaltic margins, project into the microgranodiorite, which itself contains small, strongly chilled basaltic pillows and pillow fragments ( Plate 3.5; Fig. 3.2

**Plate 3.1.** Quench plagioclase laths in microgranophyric groundmass, M1 microgranodiorite, Glendrian Bay. Note that the patches of granophyre around the feldspar laths are separated from each other by areas of granular - textured altered groundmass, in a similar arrangement to that characteristic of devitrified felsic rocks elsewhere in the aureole and contact zone of the MBG ( Plates 3.3 and 3.8 in particular ). This is consistent with the interpretation of granophyric texture as the product of coarsening of spherulitic texture ( Lofgren 1971 ). Sample 46A2. Crossed polars, field of view 1.5mm.

**Plate 3.2.** Patches of spherulitic devitrified glass in a largely holocrystalline M1 microgranodiorite, Glendrian Bay. Sample 24/1. Plane polarised light, field of view 3.0mm.



**Plate 3.3.** Devitrified glass spherulites nucleated on quench plagioclase crystals, M1 microgranodiorite, Glendrian Bay. Granular - textured areas of quartz + alkali feldspar + chlorite  $\pm$  epidote between the spherules may have formed by replacement of glass which had not yet devitrified when alteration began. Sample 24/1. Crossed polars, field of view 1.5mm.



**Plate 3.4.** Lobate pillowed contact between doleritic MBG basic rock and felsic to mesocratic hybrid rocks, contact zone of the MBG, south shore of Glendrian Bay. Grid reference 45857030.

**Plate 3.5.** Chilled and fragmented fine - grained basic pillows in M1 microgranodiorite, south side of Glendrian Bay. Grid reference 45837031.

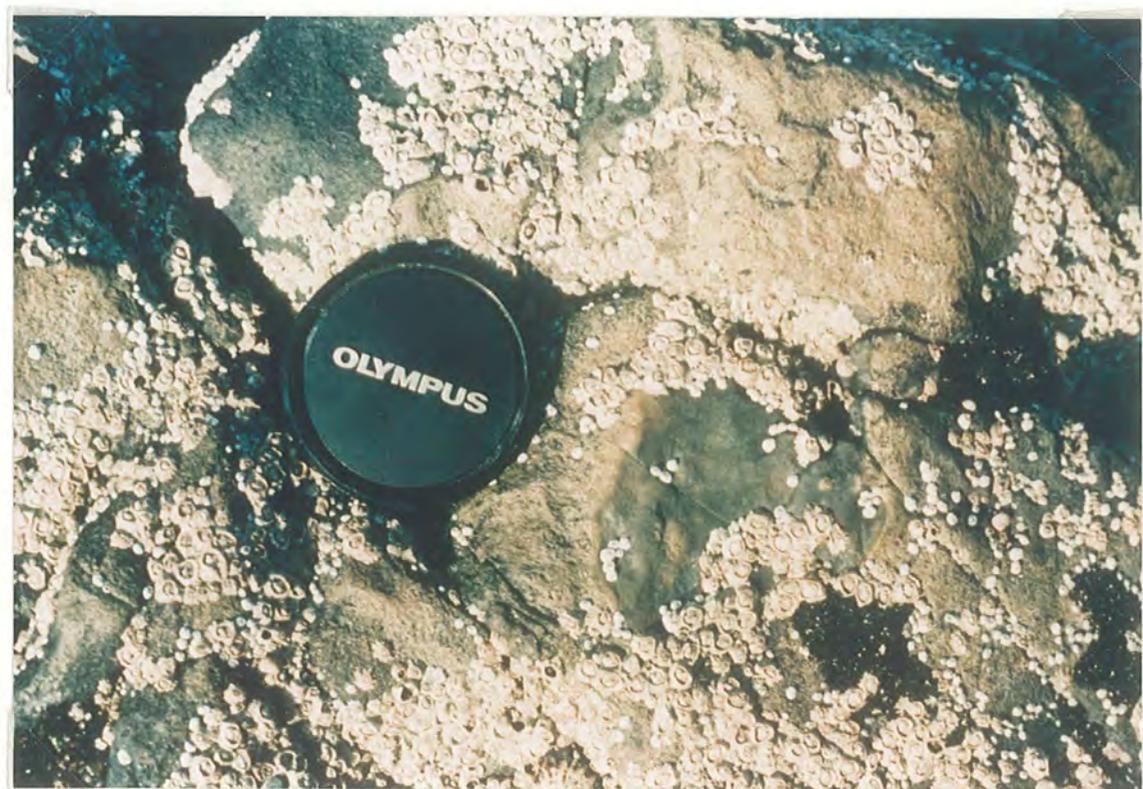
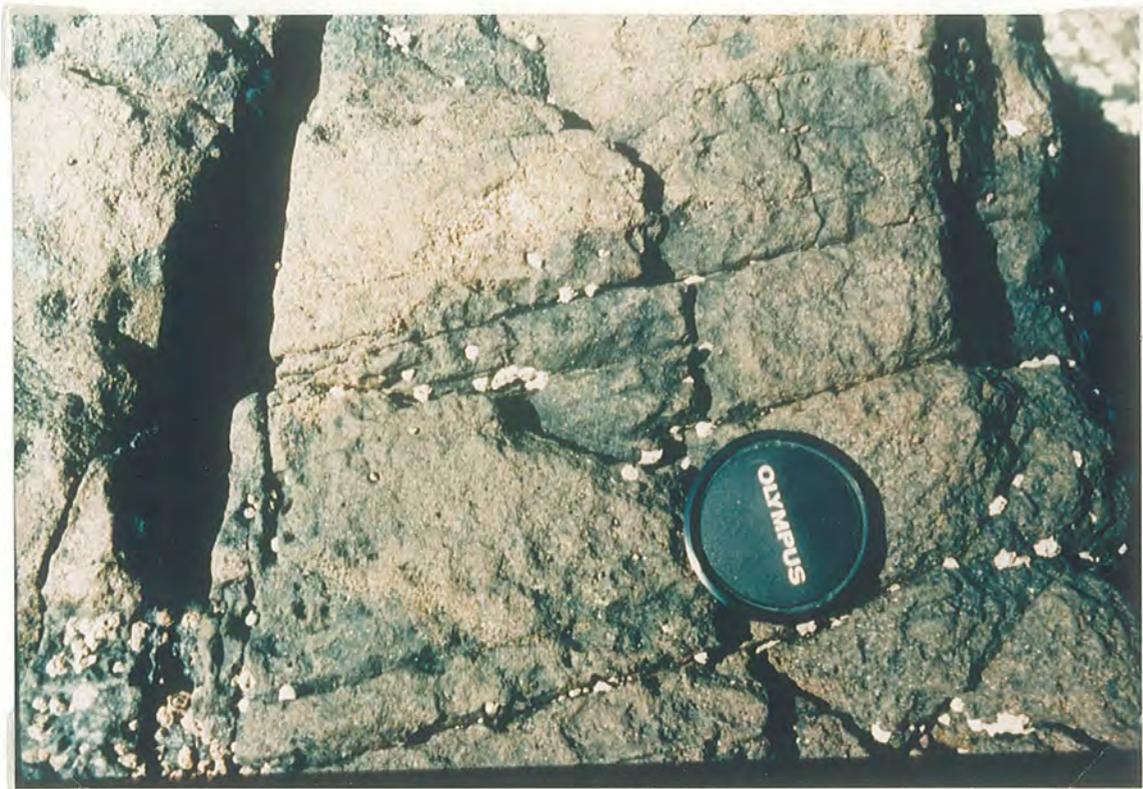


Fig. 3.2. Field sketches illustrating contacts between basic rocks and heterogenous hybrid rocks (A) or microgranodiorites (B) which co - existed as magmas prior to quenching of the MBG contact zone in the Glendrian Bay area.

A.

Heterogenous hybrid rocks with decimetre - scale mottling caused by compositional and grain - size variation

Lobate or cauliform chilled margin

Diffuse veins cutting chilled margin.

Near - aphyric basaltic to doleritic rock.

0 15cm

B. Basic pillows and pillow fragments in microgranodiorite.

Small diffuse medium - grained inclusions in microgranodiorite; probably cognate (see section 4.2.2).

Cracked pillow rims or "rinds"

Veins cutting chilled margin

0 10cm

). However, where heterogenous hybrid rocks are in contact with the microgranodiorite chilling of the former is much less strongly developed, although the contact remains sharp and lobate ( in the area around 46017207, for example ). The presence of lobate and chilled margins of this type is important because it indicates that **the microgranodioritic, heterogenous hybrid and basic rocks all formed from magmas of contrasting temperatures and crystallisation temperatures which coexisted at the margin of the magma chamber represented by the Marginal Border Group rocks in this area.** The magmatic structures present in these rocks are comparable with those attributed to mixing or conmingling of magmas of different temperatures and compositions in, for example hybrid ring - dyke complexes ( Wager & Bailey 1953, Blake et al. 1965, Sparks & Marshall 1986 ).

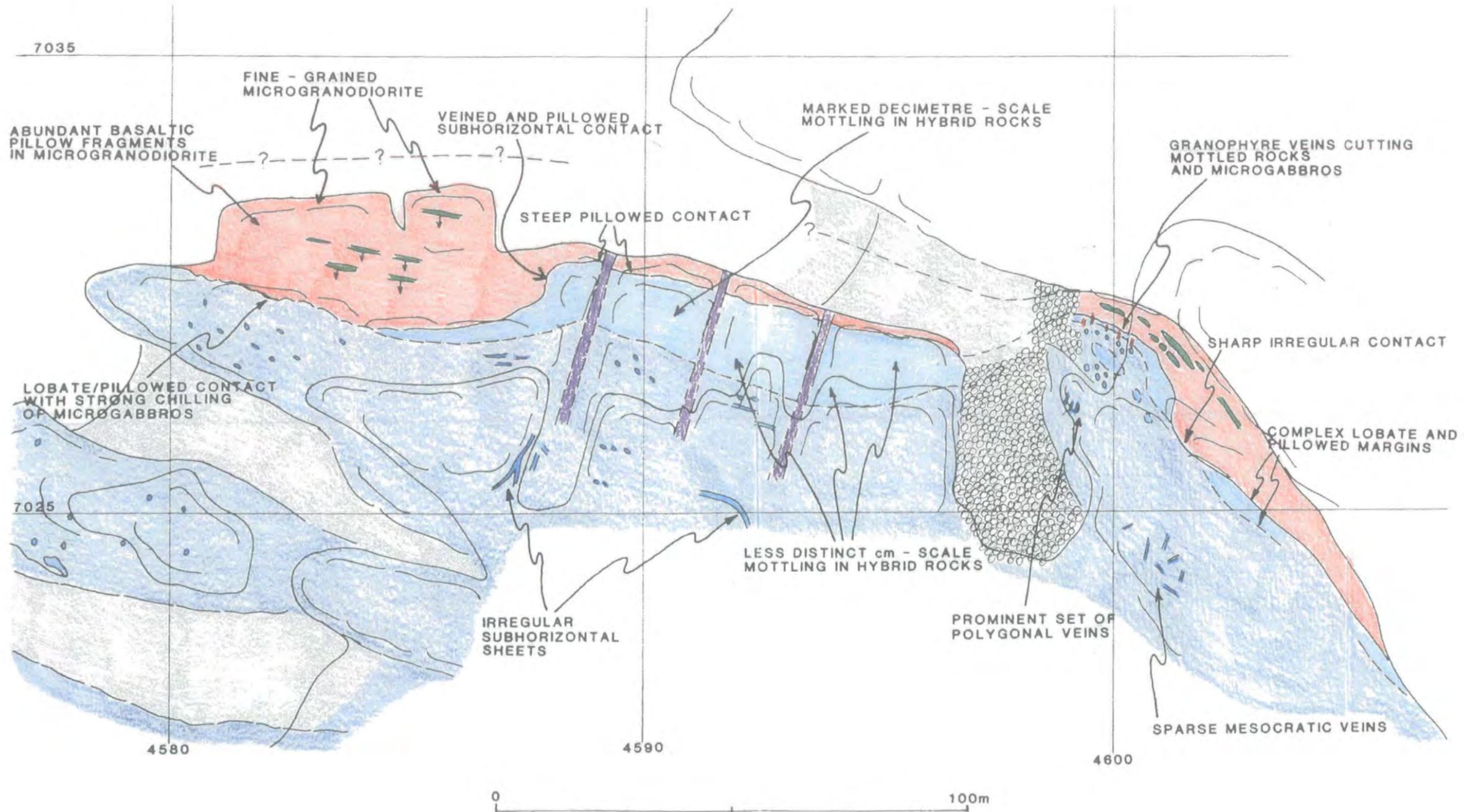
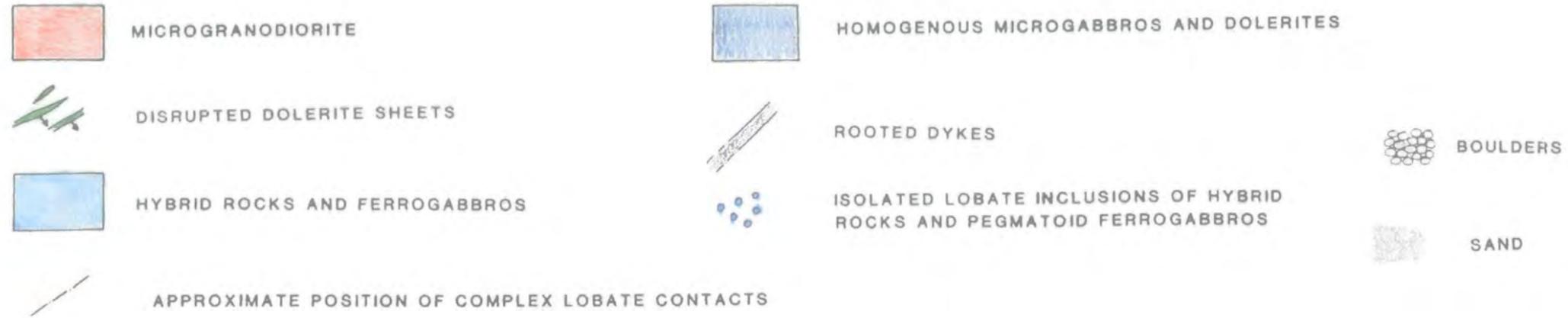
As noted above, the microgranodiorite is in places in direct contact with fine - grained dolerite and microgabbro. Elsewhere, a tract of heterogenous rocks up to 20m wide separates the two. These heterogenous rocks are very varied both in composition and grain size. This variation occurs both parallel and perpendicular to the contact and does not appear to show a systematic pattern. Fig. 3.3 shows the distribution of the main types of heterogenous rock, which may be divided up as follows:

1). Relatively leucocratic rocks, typically medium - grained ( although markedly coarser than the adjacent microgranodiorite ) with a faint decimetre - scale mottling defined by the contrast between slightly coarser - grained, more leucocratic rock and finer - grained, more mafic material. These rocks typically contain small chilled pillows and unchilled lobate inclusions of microgabbro which appear to represent a later mingling of mafic magma with a much cooler and more silicic heterogenous hybrid magma ( Plate 3.6 ).

2). Intensely heterogenous mottled rock ( Plate 3.7 ), composed of diffuse blebs and streaks of coarse, sometimes pegmatoid - textured mesocratic to leucocratic rock in a finer - grained mafic host. There is a crude correlation between the scale of the heterogeneity and its intensity, with the contrast between the two components of the rock being much weaker in rocks which show centimetre - scale mottling than that in Plate 3.7.

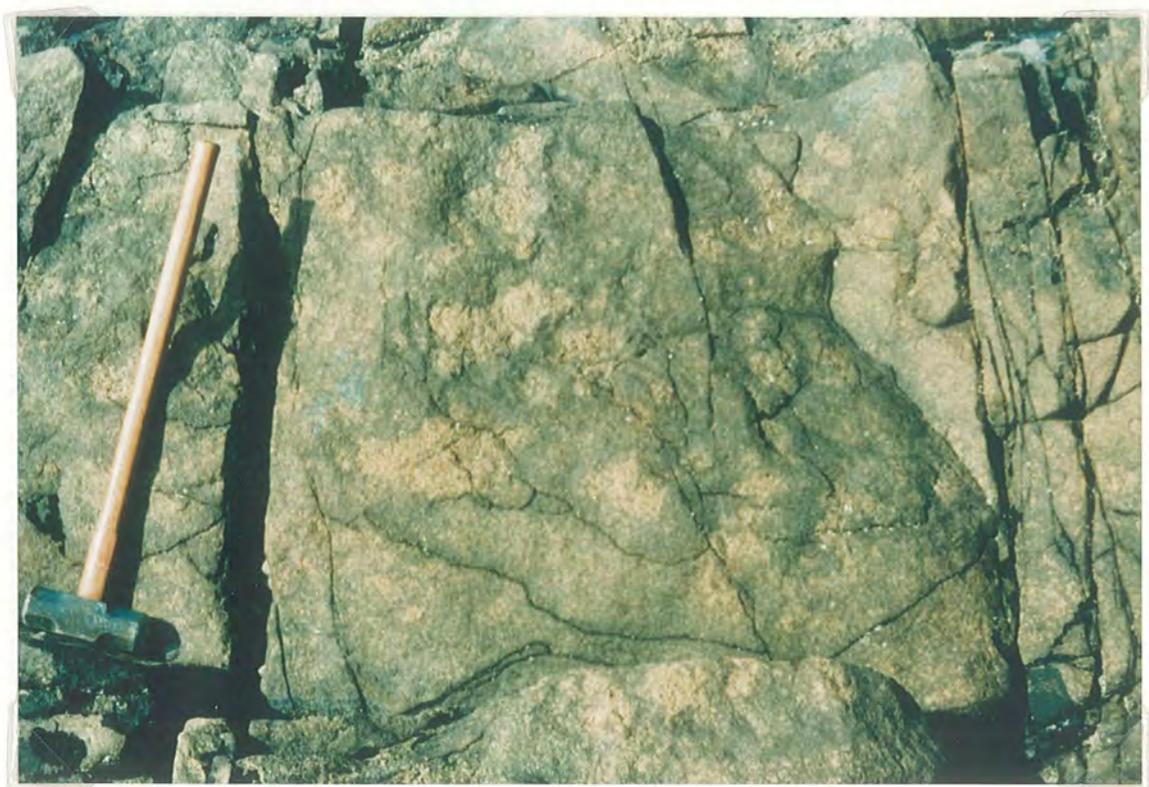
3). Largely homogenous isotropic microgabbros containing sparsely distributed, diffuse lobate inclusions. These are composed of coarse - grained, sometimes pegmatoid mesocratic rock which in

FIG. 3.3. SKETCH MAP SHOWING CONTACT ZONE OF THE MARGINAL BORDER GROUP ON THE SOUTH SIDE OF GLENDRIAN BAY. ( LATER DYKES OMITTED )



**Plate 3.6.** Lobate, possibly chilled microgabbroic inclusion in faintly heterogenous hybrid rock, south east side of Glendrian Bay. Grid reference 46017027.

**Plate 3.7.** Strongly heterogenous mottled hybrid rock, south shore of Glendrian Bay. The pale areas are dioritic to granitic in composition, the dark areas are more basic and finer - grained. Grid reference 45907029.



some is unusually rich in magnetite.

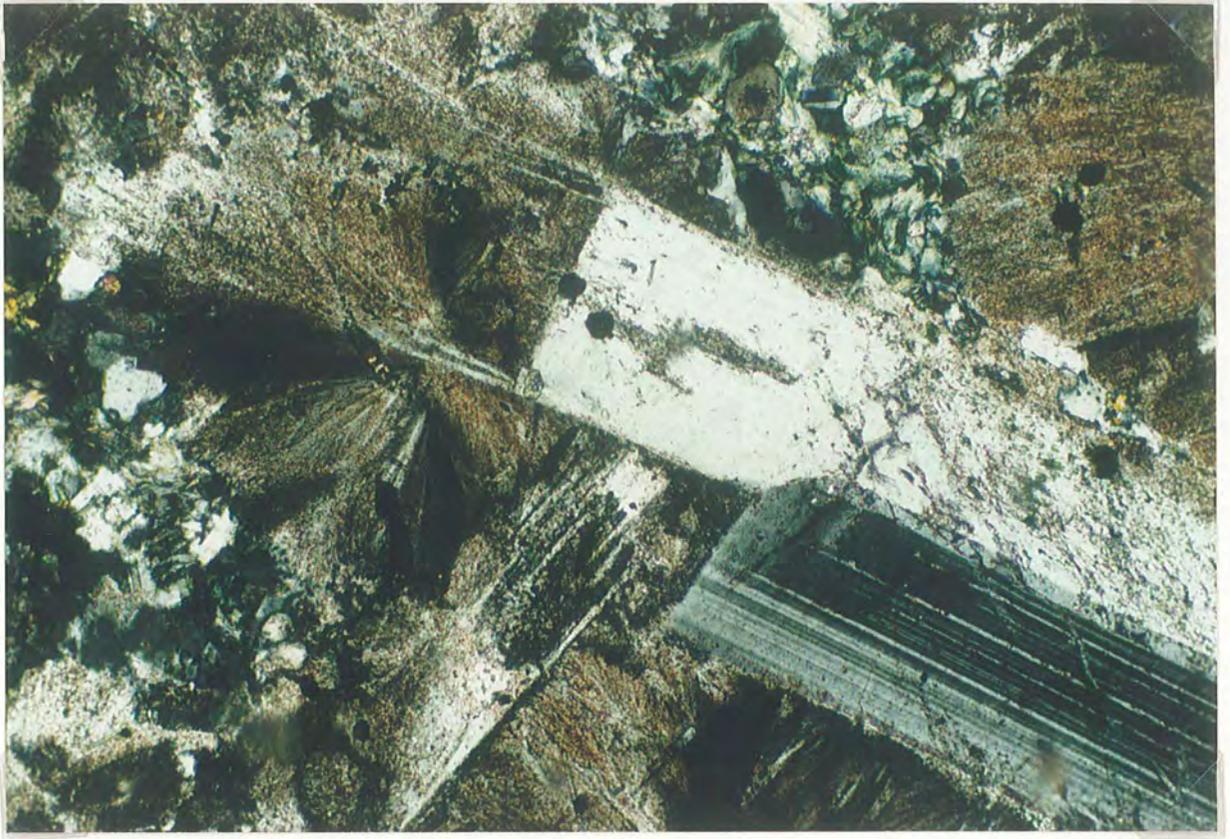
The contacts between occurrences of these three rock types are often sharp, although there is much variation within the categories defined above. Contacts between the first two and the homogenous microgabbros are often lobate, but chilling of the mafic rocks is only weakly developed.

The heterogenous nature of the mottled rocks is also evident in thin section: diffuse glomerocrysts of strongly zoned calcic feldspar and subophitic augite occur in a groundmass of zoned euhedral feldspar laths, together with augite and opaques, enclosed in interstitial microgranophyre and/or devitrified glass. The ratio of euhedral crystals to silicic interstitial material varies widely on a millimetre to centimetre scale within the sections, reflecting compositional variation from near - basaltic to granodioritic compositions on this scale. The presence of microgranophyre, spherulitic textures and quench overgrowths on feldspar laths ( Plate 3.8 ) indicates that these rocks also underwent a two - stage cooling process involving initial slow cooling and subsequent very rapid cooling, resulting in crystallisation at high degrees of supercooling and the eventual formation of residual glasses. As with the microgranodiorite, devitrification of the latter was halted by replacement of the remaining glass by hydrothermal assemblages ( epidote + quartz + chlorite  $\pm$  adularia ). The presence of magma - mixing structures implies that the rocks crystallised entirely in situ and that, as with the microgranodiorite, the two - stage crystallisation process also developed in situ.

Overall, the compositional and textural heterogeneity exhibited by these rocks suggests that they formed by incomplete magma mixing, or magma mingling ( Sparks & Marshall 1986 ). They do not have a discrete intrusive contact with the homogenous microgabbros that form the bulk of the MBG in this area and it therefore appears that they formed at the wall of the magma chamber represented by the microgabbros. The nature and origin of the magmas involved in the mixing process will be discussed further in Chapter 4. However, two points which are apparent from field observations and petrography alone are worth emphasising at this stage. Firstly, a wide range of magmas was involved in the mixing process. Secondly, there is little evidence for the large - scale involvement of melts derived from the immediately adjacent wall rocks, although a few small metasedimentary xenoliths do occur: a single dark green, granular ferruginous xenolith was found at 45847030 and Wells ( 1951 ) describes aegirine - bearing quartzose xenoliths from this area, although these were not found during the course of this work.

**Plate 3.8.** Plagioclase laths in patch of quenched and devitrified felsic groundmass within rooted dolerite dyke, Glendrian Bay. The interstices between the spherules are filled with granular quartz + chlorite + feldspar + epidote ( c.f. Plate 3.3 ). Note quench overgrowths on otherwise euhedral ( although hollow ) plagioclase lath in centre of field of view: these are comparable to overgrowths produced experimentally by increasing the rate of cooling during crystallisation experiments ( see Lofgren ( 1980 ), especially Fig. 9c, p.502 ). Sample 24/3. Crossed polars, field of view 1.5mm.

**Plate 3.9.** Cluster of four plagioclase laths radiating from ( and nucleated upon ? ) edge of a larger plagioclase crystal ( at top of plate ). Doleritic - textured homogenous MBG microgabbro, south side of Glendrian Bay. Sample 193/2. Crossed polars, field of view 0.75mm.



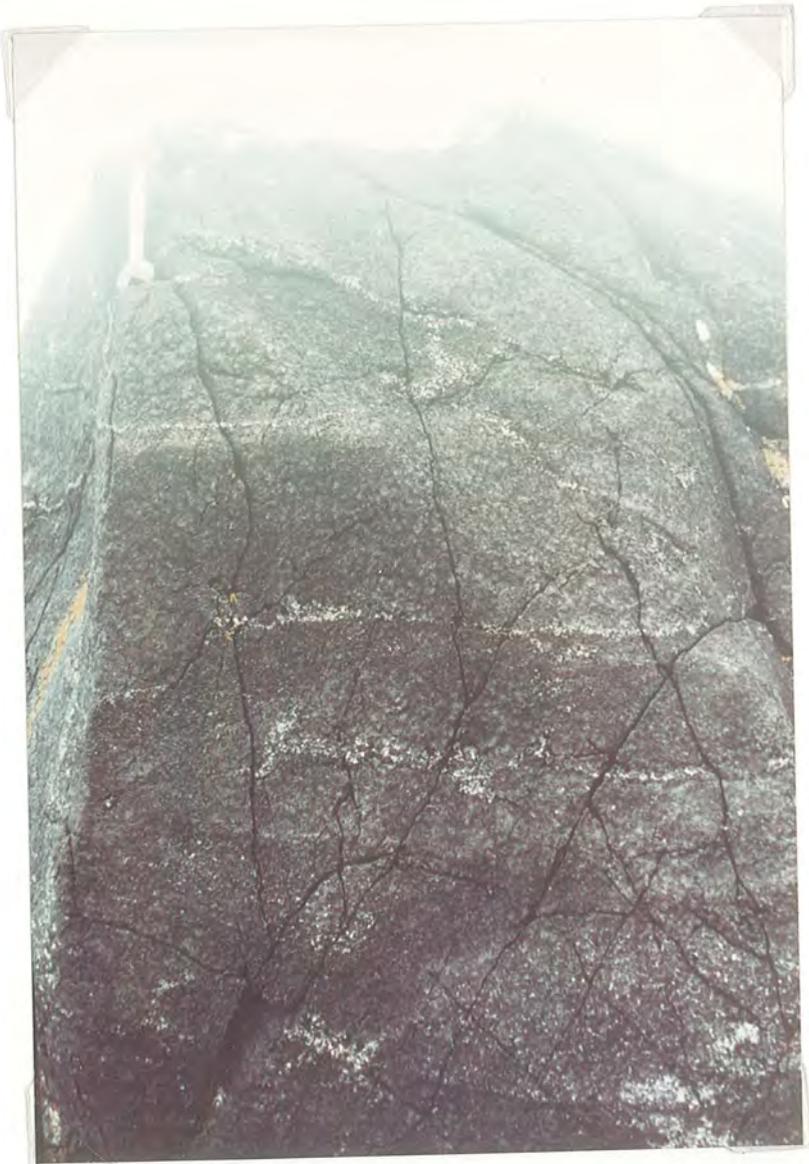
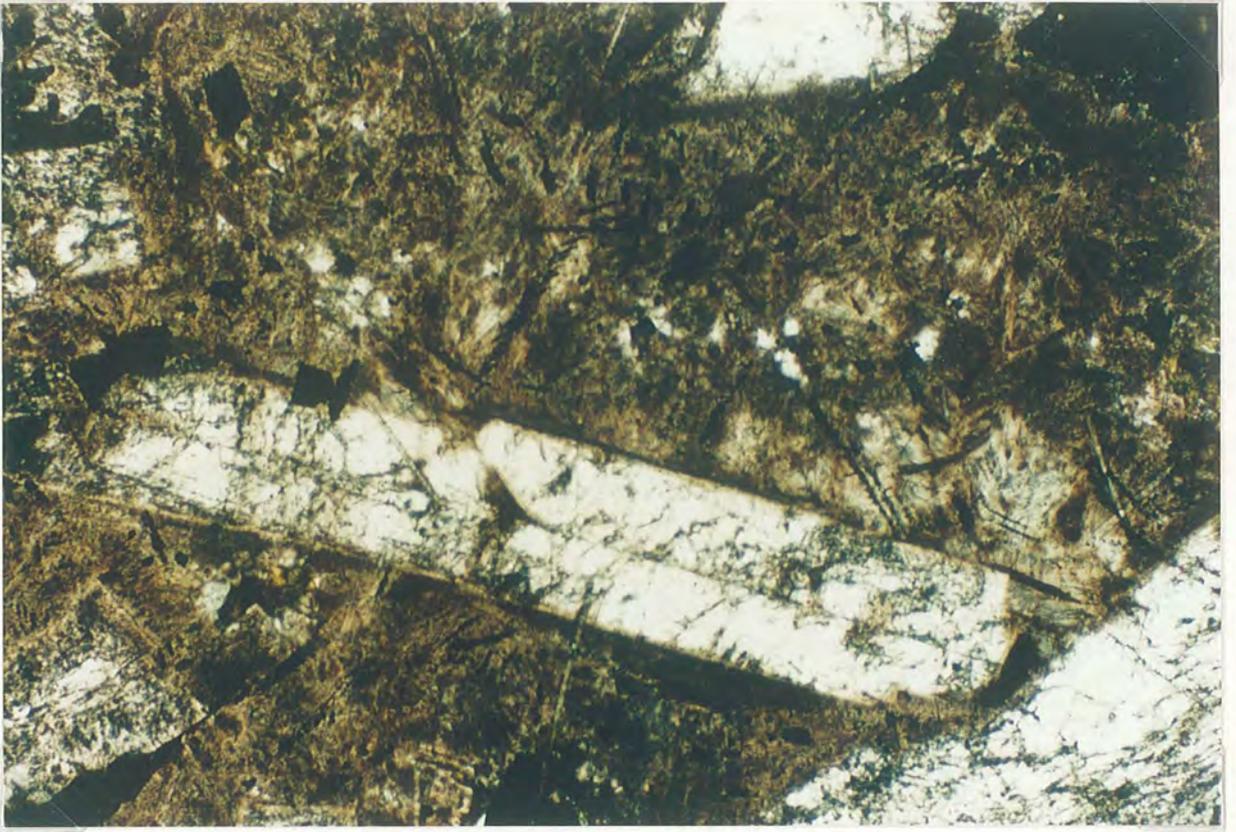
Inboard of the zone of mingled - magma rocks, the fine- to medium - grained homogenous isotropic microgabbros coarsen gradually away from the contact over a distance of up to a few tens of metres. They are typically dark aphyric rocks, commonly rather altered but retaining near - doleritic textures with prominent ophitic pyroxenes. In thin section, quartz is very rare or absent, altered olivine is sometimes present and orthopyroxene is an ubiquitous accessory phase. Radiate and sheaf - like clusters of elongate plagioclase laths with intergrown augite are present ( Plate 3.9 ), suggesting heterogenous nucleation and crystallisation under moderate degrees of supercooling. The rocks are in general unusually fine - grained for the margins of such a large pluton, as was originally noted by Richey et al. ( 1930 ).

The microgabbros contain a variety of veins and inclusions. Away from the contact, swarms of diffuse lobate inclusions are still common. These are, however, all composed of very coarse, pegmatoid - textured, magnetite - rich gabbro. The distribution of these inclusions is very uneven: almost absent to the east of Easting 459, they are very common in grid squares 457702 and 458702. Within the swarms, which are up to a few tens of metres long, they form 5% or more of the rock. Like most of the other more evolved rocks in the Glendrian Bay area, the inclusions contain interstitial devitrified glass, which in this case contains abundant quench opaque and apatite needles and is relatively poor in quartz ( Plate 3.10 ). This suggests that these glasses were less fractionated than those found closer to the contact, which implies in turn that they were quenched at both a higher temperature and at a higher cooling rate ( Lofgren 1980 ).

Closer to the contact, the microgabbros are cut by veins and irregular unchilled sheets of coarse leucogabbro, together with a few veins of mottled hybrid rock. These are usually subhorizontal but at 45947027 ( Field Loc. 45 ) a vertical outcrop forming two sides of a water - worn crag contains a network of joint - like gabbroic veins which resemble a columnar joint network, with the long axes of the columns perpendicular to the contact ( Plate 3.11 ). This structure may represent a network of thermal contraction cracks produced by progressive cooling from the contact inwards, analogous to columnar jointing in lava flows, which was filled by residual or externally - supplied fractionated melts. The subhorizontal sheets are larger, up to 20cm thick and laterally continuous for distances of 5m or more, although rather irregular. Their planar geometry and orientation suggests that they were produced by deviatoric rather than thermal stresses. Small, irregular felsite veins cut the hybrid rocks and microgabbros close to the contact but are not developed on anything like the scale

**Plate 3.10.** Plagioclase laths in devitrified interstitial glass, coarse - grained ferrogabbroic inclusion in MBG microgabbro, south shore of Glendrian Bay. Sample 193/1. Plane polarised light, field of view 3.0mm.

**Plate 3.11.** Diffuse segregation veins forming a crude columnar joint network in MBG microgabbros, south east of Glendrian Bay. Photograph taken looking down long axes of joints toward the contact. Grid reference 45997026.



**Plate 3.12.** Granular hornfelsed dolerite sheets brecciated and veined by M1 microgranodiorite. South east shore of Glendrian Bay, grid reference 46017028.



seen at Duin Bhain ( section 3.2.3 ).

In contrast to the country rocks to the north and east, the MBG around Glendrian Bay is largely free of cone sheets. The most important exception to this rule is provided by a number of rather irregular dolerite sheets which cut the microgranodiorite ( Fig. 3.3 ). Although they have well - developed chilled margins, these sheets are extensively back - veined and disrupted by the microgranodiorite, which also contains spalled fragments of the chilled margins. The scale of the disruption ( Plate 3.12 ) and the textural uniformity of the microgranodiorite suggests that this was not simply caused by local remelting of the host by the dolerite sheets but reflects intrusion of the dolerites into a hot host which exhibited brittle or plastic behaviour according to strain rate. The dolerites themselves are partly hornfelsed to granular augite - feldspar - opaque assemblages and contain annealed microfaults. These are marked in thin section by fine - grained granoblastic crush zones, .5 to 2mm thick, with the same assemblage as in the host dolerite. This deformation is problematic because it is not present in the microgranodiorite and predates high grade, pyroxene hornfels facies metamorphism in the dolerites which is not recorded in the microgranodiorite; there is, for example, no evidence of remelting and corrosion of the primary feldspars in the latter ( except possibly in the rare feldspar porphyrocrysts ). The most obvious interpretation, that the dolerites were intruded and deformed prior to full crystallisation of the microgranodiorite host, seems mechanically unlikely because of the high stresses needed to produce cataclastic deformation in the dolerite.

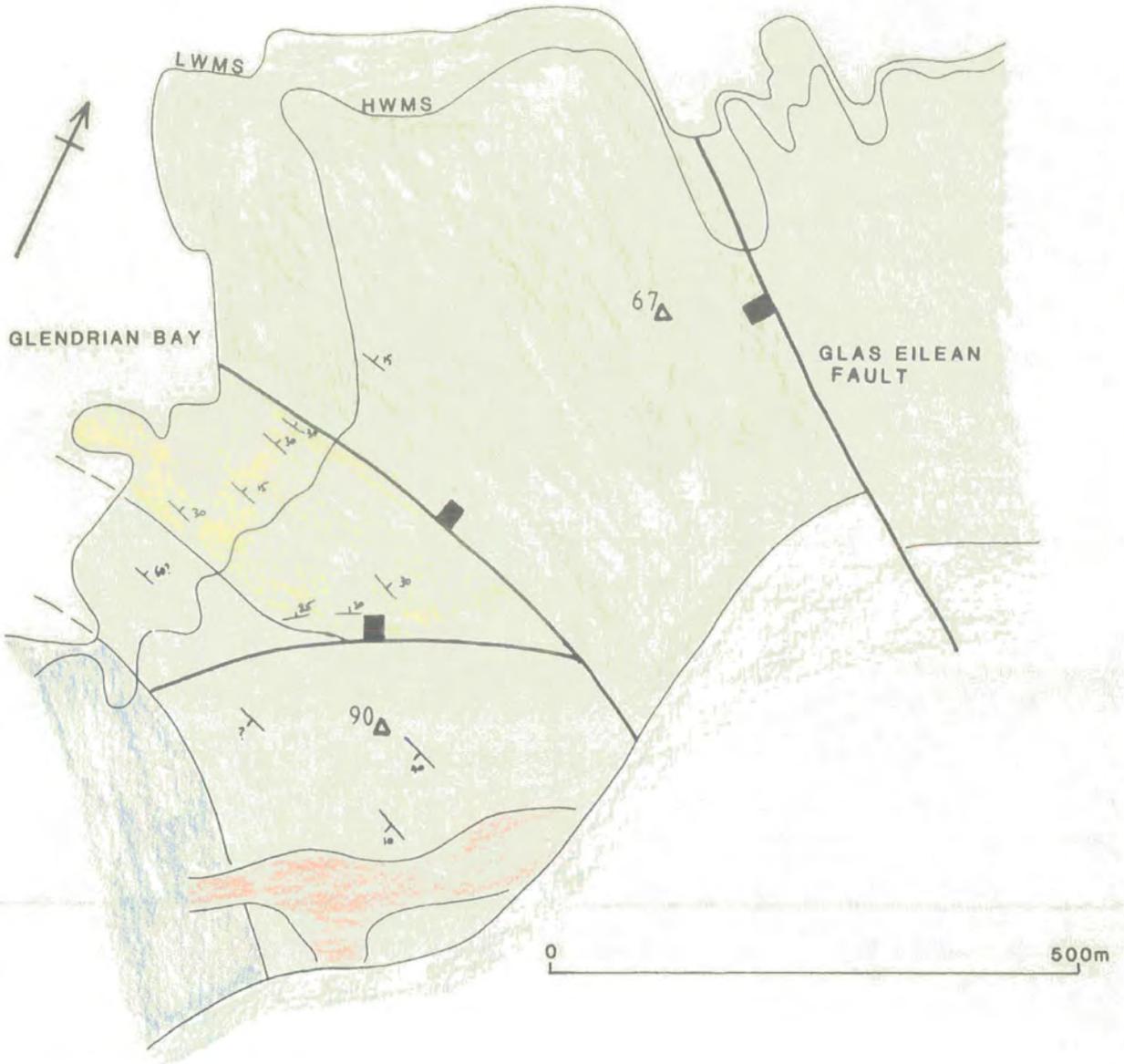
A number of dykes and cone sheets cut the coarser rocks of the MBG but most are very fresh, very fine - grained and have prominent chilled margins. These are very much younger than the MBG and will not be considered further. However, a number of dykes around 45927029 form a distinct early group. These are almost aphyric, with only a few plagioclase phenocrysts, coarser - grained and lack chilled margins except where they cut the microgranodiorite. To the south of the microgranodiorite - hybrid rock contact they lose their chilled margins, become coarser and less regular, and finally grade into the host microgabbros. This rapid spatial change in character implies that these dykes were emplaced when the country rocks were cool but the magma chamber margin was still within a few tens of metres of the original contact, soon after the rapid quenching of the country rocks ( see below ) and of the microgranodiorite. In the sense that they can be traced back into their source, these dykes can be considered to be rooted dykes ( Pedersen 1986 ).

### 3.2.2.2. The country rocks in the Glendrian Bay area, and Minor Intrusions associated with the Pluton.

To the north and east of the contact, as far as the edge of the Tertiary volcanoclastic rocks which form the promontory of Rubha Carrach, are a suite of rocks which have previously been mapped as hornfelsed Jurassic sediments and pre - Hypersthene Gabbro cone sheets of the 'outer cone sheets of Centre 2' ( Richey et al. 1930 ) and as metasomatised sediments and cone sheets ( Wells 1951, 1954 ). Remapping of this area ( Map 2 ) has shown these earlier works to be incorrect. Four groups of rocks have been identified. In order of decreasing age, these are: hornfelsed Jurassic sediments, rheomorphic breccias ( corresponding to Wells' metasomatic pseudobreccias ), felsite intrusions and dolerite cone sheets.

As noted in Chapter 2, the Jurassic rocks in the Glendrian Bay area probably belong to the Lower Liassic Pabba Beds and the Middle Liassic Scalpa Sandstone. Close to the shore, the rocks young to the northeast, and the few dip measurements possible also indicate overall northeasterly dip, decreasing from  $30^\circ$  near the contact to about  $15^\circ$  in the Tertiary rocks to the northeast. On Hill 90, south of an east - west fault ( Fig. 3.4 ) the structure is more complex and calc - pelitic hornfelsed occur to the southwest of underlying pelitic hornfelsed: these probably correspond, respectively, to the upper and middle units of the Pabba Beds ( see Chapter 2 ). All of these rocks are intensely metamorphosed and show evidence of at least two phases of metamorphism, M1 and M2. The first of these is of very high grade close to the contact with the Hypersthene Gabbro: hypersthene - bearing assemblages occur in Ca - poor ferruginous metapelites ( typically quartz + feldspar + opaques + hypersthene  $\pm$  cordierite ) whilst the more Ca - rich protoliths were metamorphosed to augite + plagioclase + quartz. Quartz - alkali feldspar intergrowths in some pelitic rocks may correspond to interstitial partial melts formed at this stage. Parallel trains of opaque grains in some of these pelites may represent relicts of a pre - M1 biotite - bearing assemblage comparable to the pre - Hypersthene Gabbro ( M0 ) 2 - mica assemblage found in pelites on the southern margin of the intrusion ( section 3.2.5.5 ). Pristine M0 assemblages are not found in the Glendrian Bay area except perhaps in the Tertiary volcanoclastic rocks to the north - east. The grey quartzose xenoliths in the microgranodiorite ( see above ) contain quench plagioclase and tridymite needles ( now pseudomorphed by quartz ) in a groundmass composed of quartz - feldspar intergrowths ( quenched melt? ) with only a small proportion of corroded residual quartz grains. Further away

FIG. 3.4. MAJOR FAULTS IN THE GLENDRIAN BAY AREA: CONE SHEETS, FELSITES, RHEOMORPHIC BRECCIAS AND OTHER MINOR INTRUSIONS OMITTED FOR CLARITY



GREAT EUCLITE



PILLOWED DOLERITE - GRANOPHYRE HYBRID INTRUSION



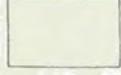
HYPERSTHENE GABBRO



TERTIARY VOLCANICLASTICS



SCALPA SANDSTONE



PABBA BEDS



FAULTS - TICK ON DOWNTOWN SIDE



DIPS IN SEDIMENTS



SPOT HEIGHTS (m)

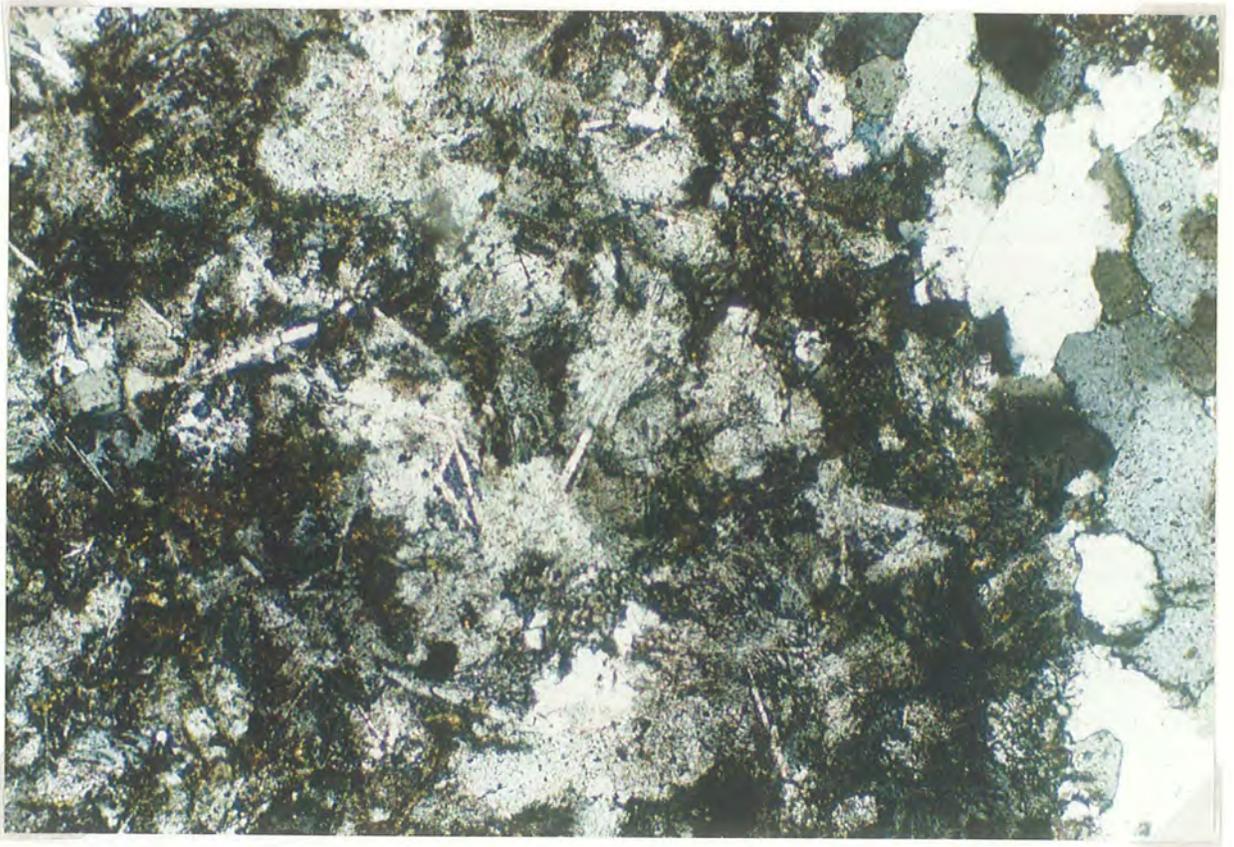
from the contact, the grade of this early metamorphism is lower, although diopside is common in the Scalpa Sandstone beds. The decline in metamorphic grade is such that pervasive hydrothermal metamorphic assemblages are preserved in the Tertiary volcanoclastic rocks to the north - east, as well as delicate primary igneous textures, such as variolitic and felsitic textures, in the clasts. The hydrothermal minerals are mainly carbonate and epidote: some later diopside is present and this may represent the outer edge of purely thermal metamorphism due to the Hypersthene Gabbro in this area. M1 metamorphism is, with one possible exception discussed below, only directly observed in the metasediments in this area. It is believed to be associated with the Hypersthene Gabbro because of the steep increase in metamorphic grade towards the latter and because some of the deformation associated with the emplacement of the pluton ( see below ) predates it. This high grade metamorphism is partially overprinted throughout the Glendrian Bay area by later, lower grade metamorphism ( M2 ) characterised by the development of hydrous minerals ( see Chapter 5 ), particularly along joints and fractures. Within 100m or so of the Great Eucrite the Jurassic rocks, along with all others in the area including the much later dolerite - granophyre mixed - magma intrusion on the south side of Hill 90, are overprinted again by more pervasive metamorphism and recrystallisation, presumably produced by the Great Eucrite. The rocks in the coastal area appear to have been unaffected by this.

The rheomorphic breccias occur over a large area of the rock platform around Glendrian Bay and also in a number of inclined sheet - like to lensoid bodies on the eastern side of Hill 90. Superficially, they resemble massive, unstructured matrix - supported breccias with a monomict assemblage of poorly sorted and randomly orientated quartzite clasts ( Plate 3.13 ). In thin section, however, the matrix shows a variety of fine - grained but invariably igneous textures. These range from granular - poikiloblastic and microgranophyric quartz - feldspar intergrowths through felsitic to spherulitic ( devitrified glass ) textures. Quench crystals are commonly present, as are corroded residual crystals and polycrystalline 'microclasts' . Quench tridymite needles pseudomorphed by quartz are ubiquitous, as is residual quartz, with a characteristic habit of sutured and strained equant grains ( Plate 3.14 ). The rest of the residual and quench crystal assemblages vary with rock composition ( see Chapters 4 and 5 ) but include plagioclase, hypersthene, diopside and opaques. Zircon is a common rounded residual accessory phase.

The quartzite clasts, actually blocks of residual material, also show features indicative of an anatexis

**Plate 3.13.** Rheomorphic breccia, south east shore of Glendrian Bay ( Field location 47A, grid reference 46007029 ). The residual quartzite blocks are the rounded, pale yellowish - grey bodies. Black patches on the rocks are surficial tarry deposits.

**Plate 3.14.** Matrix of rheomorphic breccia. Note residual quartz microclast at extreme right, inverted tridymite needles pseudomorphed by quartz ( centre of field of view ) and felsitic - textured quartz + feldspar rich groundmass. Sample 47C2. Crossed polars, field of view 1.5mm.



origin. Apart from a few pure quartz blocks, they have a spongy texture defined by rounded pockets of grey material similar to the matrix when seen at outcrop ( Plate 3.15 ). In thin section these pores are composed of zoned quartz - feldspar intergrowths, with an outer zone rich in quartz and containing quartz pseudomorphs after quench tridymite, and an inner zone composed of a single rounded feldspar grain intergrown with a number of poikilitic quartz grains ( Plates 3.16, 3.17 ). The development of these textures is discussed further in Chapters 4 and 5 but it is apparent that they were produced by melting and crystallisation at the sites of feldspar grains in the protolith.

The origin of these rocks is discussed in detail in section 4.2.4. However, the outcrop and petrographic evidence alone suggests that they were produced by partial melting of a bedded sedimentary sequence composed of feldspathic sandstones interbedded with less quartz - rich sediments. The latter must have been more fusible as they underwent almost complete melting, to produce at least part of the matrix ( see section 4.2.4 ), whilst the sandstones, although partially melted, remained coherent. Subsequent disruption of the sandstone beds by the melted material, partly by dissolution in the melt but perhaps also by physical displacement, could then have produced the breccia - like appearance of these rocks. Similar rheomorphic breccias have been described from the contact zones of other upper crustal mafic intrusions, notably the type examples around mafic intrusions in the Monteregian province of eastern Canada ( Philpotts 1970 ), and those around the Rhum ultrabasic complex ( Greenwood 1987 ). Unlike these examples, however, the rheomorphic breccias in the Glendrian Bay area appear to have been intruded into the surrounding country rocks on a large scale. On the eastern side of Hill 90 in particular, the breccias occur in sheetlike and lensoid bodies, dipping towards the Hypersthene Gabbro, which are discordant to the host Jurassic metasediments. These bodies resemble cone sheets in their overall geometry.

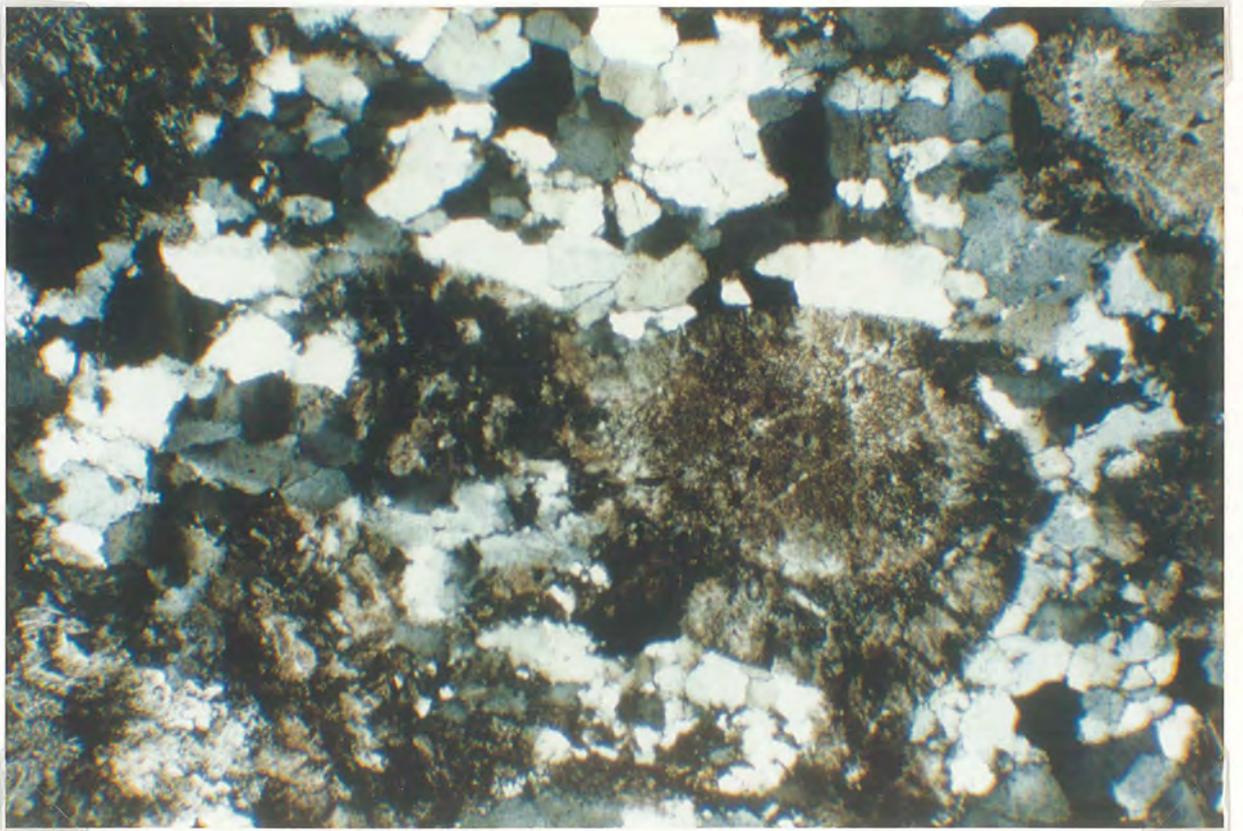
As noted above the rheomorphic breccias contain quench textures indicative of very rapid cooling, like the microgranites and other rocks in the MBG. Metamorphic overprinting of these textures is restricted to devitrification and hydration of the groundmass and hydrous alteration of the quench and residual crystals, except in the aureole of the Great Eucrite. There is no evidence in the breccias for high - grade metamorphism of the type seen in the Jurassic rocks. It would appear that the rheomorphic breccias were formed and intruded during M1, quenched at its end and then affected by the subsequent hydrothermal alteration. This would be consistent with the age relationship of the breccias to the microgranodiorite at the contact with the Hypersthene Gabbro ( see above ). One

**Plate 3.15.** 'Sponge' texture in residual quartzite blocks, produced by the presence of rounded pockets of quenched melt. Location 47A, grid reference 46007029.



**Plate 3.16.** 'Sponge' textured residual quartzite block in thin section, sample 47C2. The dark round patches in the centres of some of the quenched melt pockets are single feldspar crystals intergrown with micropoikiloblastic quartz crystals ( **Plate 3.17** ). Plane polarised light, field of view 14mm.

**Plate 3.17.** Detail of quenched melt pocket in sample 47C2, photographed under crossed polars. Rounded feldspar grain at centre of pocket is fringed by bright - polarised quartz needles pseudomorphous after tridymite. Light and dark patches within the feldspar crystal are produced by the different orientations of the micropoikiloblastic quartz crystals with which it is intergrown. field of view 1.5mm.



problem posed by this interpretation is that none of the rheomorphic breccia intrusions, including those emplaced at a relatively great distance from the contact, show any sign of slow pre - quench cooling. This would be expected to occur in those rheomorphic breccia intrusions emplaced into the outer, cooler parts of the contact aureole in particular. This would be minimised if they were emplaced close to the end of M1, at which time the aureole was hottest ( see section 5.2. ) and the time available for cooling was at a minimum.

The rheomorphic breccias are the principal hosts for the numerous small felsite intrusions in the Glendrian Bay area ( Map 2 ) although these also intrude pelitic hornfelses on the eastern side of Hill 90. The felsite intrusions are irregular bodies with lensoid cross - sections, also resembling cone - sheet intrusions. They range in length from a few tens of centimetres, close to the MBG contact, up to several tens of metres ( Map 2 ). All are fine - grained, with felsitic, orbicular or spherulitic textures again indicative of quenching at high degrees of supercooling ( Lofgren 1971 ).

The felsites are invariably porphyritic, and contain two generations of plagioclase porphyrocrysts: early corroded and sieve - textured residual grains and later zoned euhedral laths. The former are typically found in glomerocrysts with augite and opaques. Chilled margins may be present in a few of the larger felsite intrusions but in general the rocks are too fine - grained overall for these to be noticeable. The felsites are not backveined by the host rheomorphic breccias and appear to have been emplaced following the abrupt solidification of the latter. Hydrothermal alteration of the felsites is common, implying that they were emplaced during the M2 hydrothermal metamorphism which followed the quenching of the rheomorphic breccias. This also implies that the felsites postdate at least part of the solidification of the Marginal Border Group: although no felsite veins were found in the MBG in the Glendrian Bay area, similar veins and sheets do cut MBG rocks south of Duin Bhain ( section 3.2.3 ).

Age relationships between the felsite intrusions and the very abundant dolerite cone sheets which also cut the rocks outside the contact are often uncertain because the two sets of intrusions are sub - parallel and cross - cutting relationships are rare. Some of the larger felsite intrusions on the eastern side of Hill 90 are, however, cut by cone sheets. As will be seen, the reverse relationship is also developed elsewhere around the contact.

Dolerite cone sheets occupy much of the outcrop in the Glendrian Bay area, particularly where cone sheet swarms are developed. These are areas, up to 100m wide in a direction radial to the contact and up to a kilometre or more long, where 80 to 100% of the outcrop is composed of cone sheets. Elsewhere, the cone sheets typically form only 20 to 40% of the outcrop. Within the swarms, individual sheets are still well defined and fine - grained, with chilled margins. The cone sheets cross - cut the Jurassic metasediments, the rheomorphic breccias and at least some of the felsite intrusions, but their age relationship to the MBG itself is rather ambiguous. Apart from the fragmented sheets in the microgranodiorite, dealt with above, neither is observed to cut the other. If the cone sheets do postdate the MBG then it seems rather odd that the dense swarms of cone sheets seen outside the contact should die out so abruptly towards it. However, this is implied indirectly by the observed age relationships between the earlier felsites and rheomorphic breccias, M1 metamorphism and the MBG. In addition only the dolerites in the microgranodiorite, and one other locality, show evidence of high grade metamorphism associated with the Hypersthene Gabbro in this area. The latter locality is at 46137017, where a cone sheet sampled only a few metres from the contact shows extensive oxidation and recrystallisation to a 2 - pyroxene hornfels assemblage. This overprints both the primary igneous textures and a series of granular crush zones and linked quartz - rich veins. The latter now contain the assemblage quartz + opaques + augite and may represent quartz - rich hydrothermal veins analogous to those associated with post - quench, M2, hydrothermal activity in the rocks exposed on the rock platform to the north. The fact that these veins have not been remelted and quenched suggests that they may not have been affected by the ultra - high - grade M1 but rather by a later episode of high - grade metamorphism present at this locality only. A further feature that distinguishes this from M1 is the presence of veins filled with orthopyroxene, opx + cpx + opaques, or quartz + opaques: the only comparable veins are those associated with hydrothermal activity in the Inner Series of the pluton. The possibility of such a localised phase of metamorphism is discussed further in Chapter 5. Elsewhere the cone sheets are affected by the post - quench M2 hydrothermal metamorphism, resulting in patchy and interstitial alteration to green hornblende and/or chlorite. Taken together, the age relationships of the cone sheets to the felsites and rheomorphic breccias, and to the metamorphism and the MBG, lead to a very important conclusion: **the cone sheets postdate the emplacement of the intrusion in its present position and therefore formed around its steep sides and not above it.** This has important implications for cone sheet emplacement mechanisms which are discussed in Chapter 7.

This conclusion is supported by the age relationships of deformation in the area. Two sets of structures can be identified. The first and older of these is a set of a few large, steeply inclined dip - slip faults which downthrow away from the Hypersthene Gabbro and predate the cone sheets. They are very poorly exposed, being largely cut out by cone sheets, but appear to be associated with the development of the outward ( doming ) dips in the Jurassic sediments and in the Tertiary volcanoclastics ( Fig. 3.4 and Chapter 7 ). The even larger Glas Eilean fault, which has a history of activity ranging from the pre - Tertiary to post - Centre 3, may also have been active at this time ( Chapter 7 ).

The second set of structures, which postdate the rheomorphic breccias, the felsite sheets and at least some of the cone sheets, are numerous but small cataclastic shear zones ( Plates 3.18 & 3.19 ). These are concentric to the contact and either dip towards it at a shallow angle or away from it, also at low angles, and appear from microstructural shear sense criteria ( Chapter 7 ) to be thrust faults, although such criteria are only observed in a few cases. Although these faults postdate the quenching of the rheomorphic breccias and the end of M1, similar faults are not found in the Marginal Border Group rocks in this area. This may, however, reflect a difference in the rheologies of the country rocks and of the MBG at the time of thrusting: as will be shown in Chapter 7, the thrusts in the country rocks and the flat - lying hybrid and ferrogabbro veins in the MBG may have formed in the same stress field. This would place the thrusting in the period also represented by the rooted dykes cutting the MBG, when the country rocks had already been chilled but the MBG was still hot and partially molten.

### **3.2.2.3. Summary of Age Relationships in the Glendrian Bay area.**

The age relationships discussed in this sub - section are summarised in Fig. 3.5. Although this applies to the Glendrian Bay area, many of the episodes and events summarised in it can be correlated with other sectors of the contact zone of the Hypersthene Gabbro. The observed age relationships are presented on a diagram of this sort rather than as a simple sequence because many of the 'events' occurred over extended and overlapping periods of time. In Fig. 3.5, solid lines denote the definite extent of a period of activity, whilst dotted lines mark the possible time - extent of activity where definite evidence is lacking. The quenching event which separates the two periods of metamorphism is marked on Fig. 3.5 as a single event. Although the mechanism proposed for the quenching

**Plate 3.18.** Cataclastic shear zone cutting rheomorphic breccias. Field location 47L, grid reference 46057037. View looks north east; shear zone dips away from viewer at up to 50°.

**Plate 3.19.** Microclasts in the very finest grained cataclastic shear zone rock from location 47L. Some clasts are composite, suggesting that they had undergone previous brecciation. There is no positive textural evidence in the groundmass to suggest that it was ever actually molten. Plane polarised light, field of view 1.5mm.

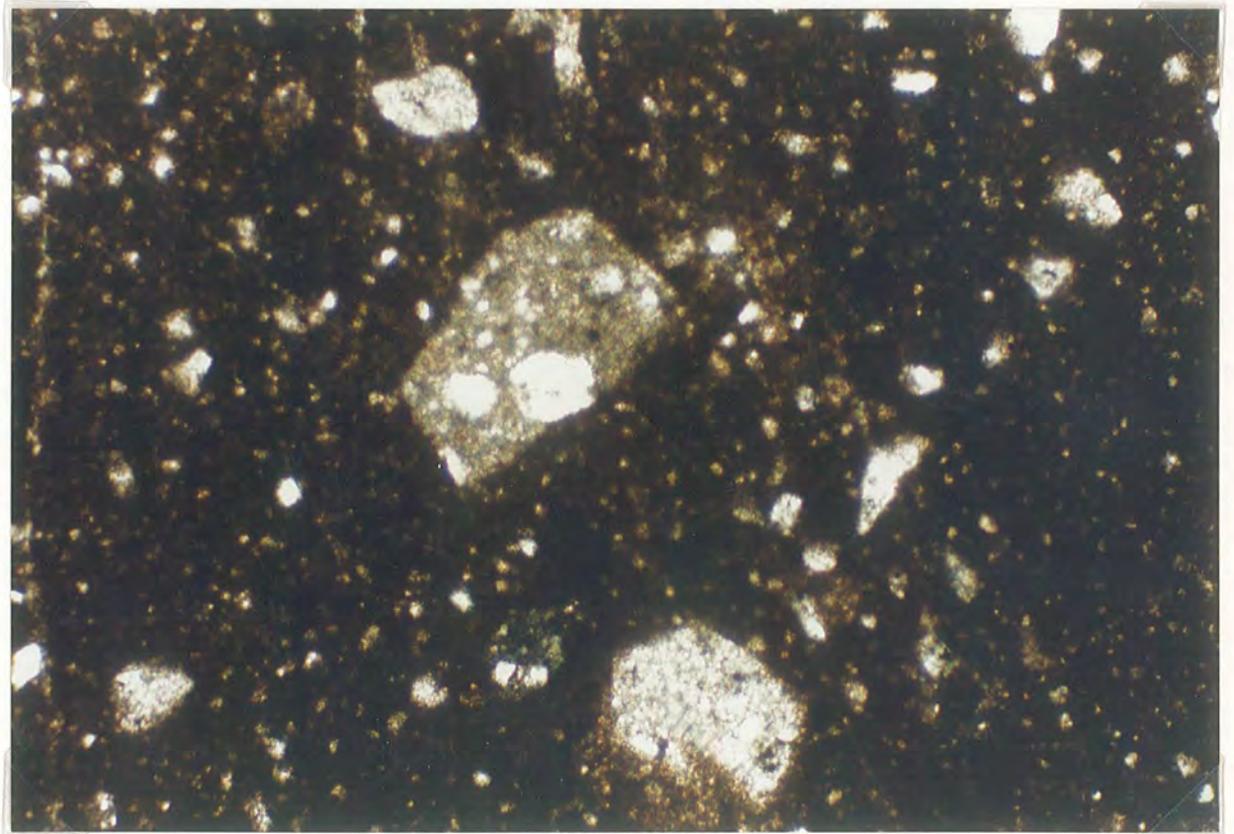
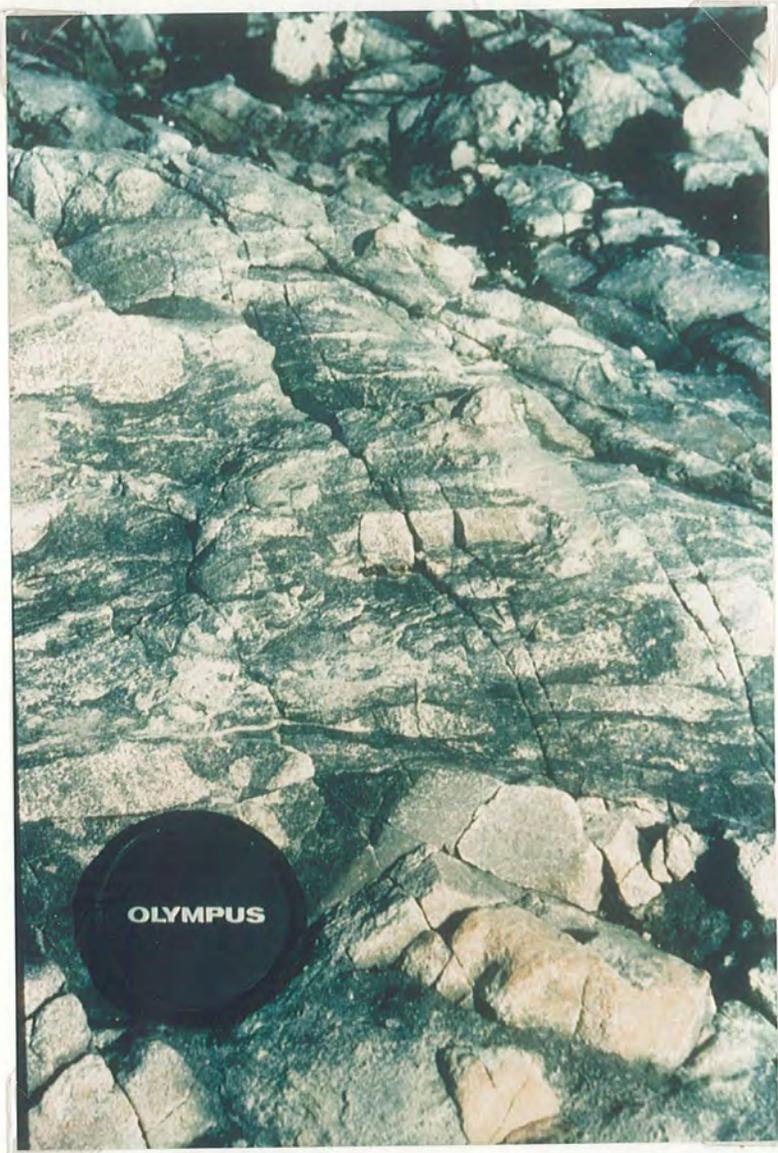
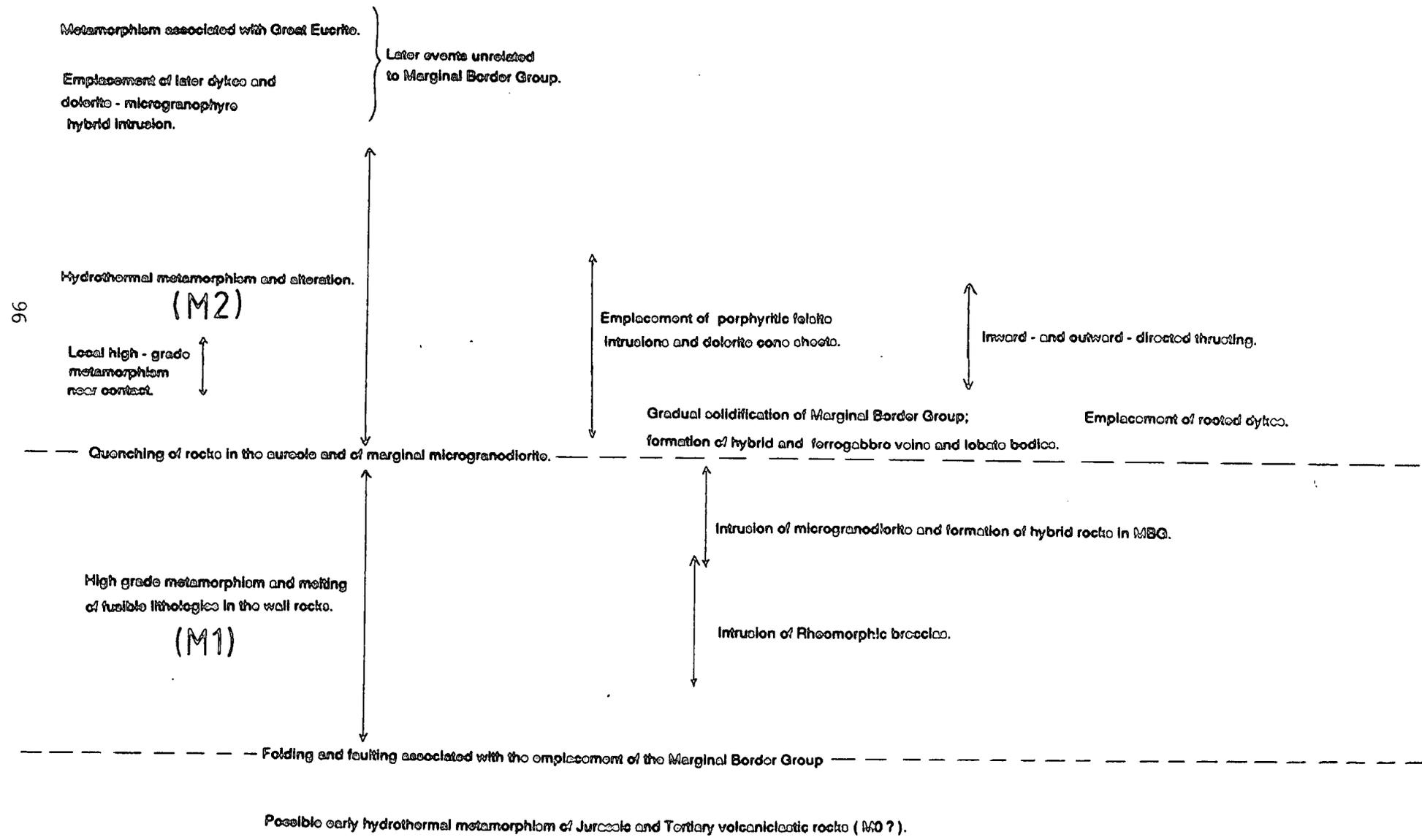


Fig. 3.5. Relative ages of intrusions, metamorphism and deformation in the Glendrian Bay sector.

DECREASING AGE



in Chapter 5 requires that it be diachronous, this diachroneity cannot be resolved except in the Marginal Border Group. The M0/M1/M2 division of the metamorphism can be applied to other events and to rocks: thus, for example, the felsites and dolerite cone sheets can be described as M2 intrusions, meaning that they were intruded during M2 metamorphism. This terminology will be applied to other areas of the contact zone.

### **3.2.3. Duin Bhain to Sanna Point ( Map 3 ).**

To the west of Glendrian Bay the outer contact of the Hypersthene Gabbro outcrops again between 45047026 and 44447038, and runs roughly parallel to the coast. West of a prominent gully in grid square 444703 it mostly lies offshore, but hybrid rocks occupy most of the coastal area as far west as Sanna Point and there are isolated occurrences of microgranodiorite and pelitic hornfelses on the northern edge of the intertidal zone.

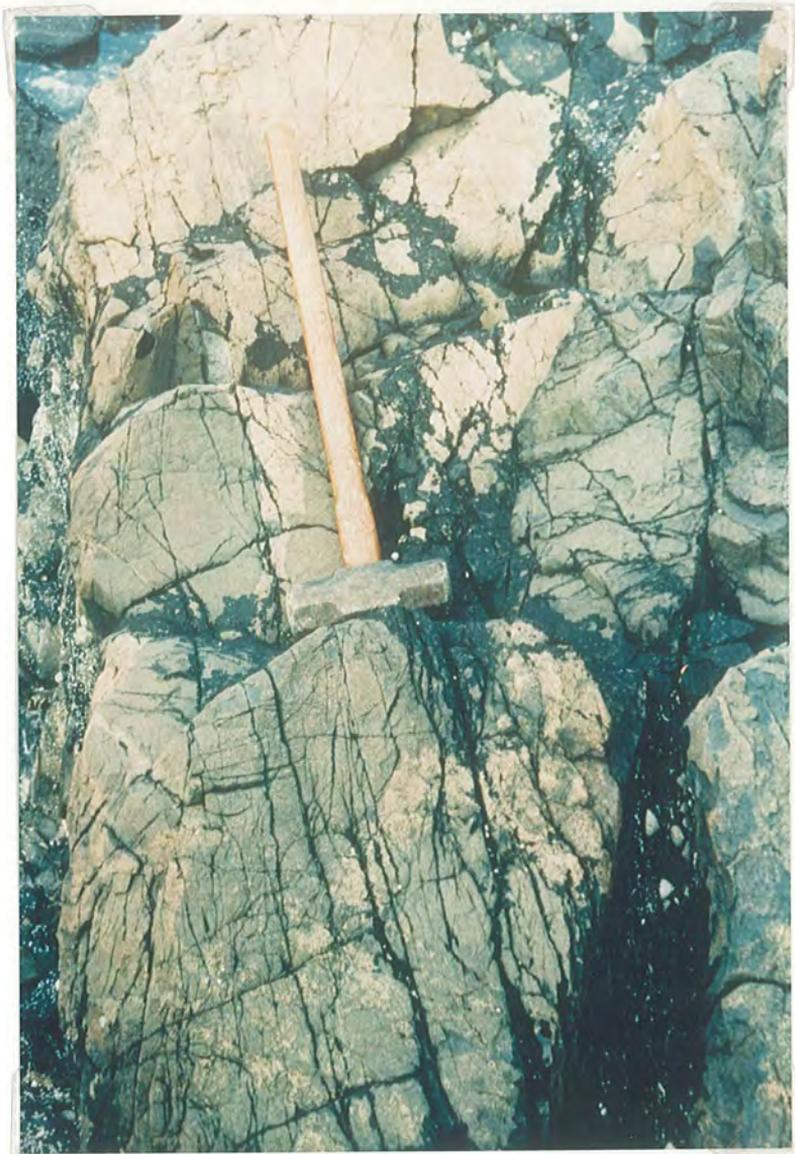
#### **3.2.3.1. M1 rocks and structures in the Contact Zone of the MBG between Duin Bhain and Sanna Point.**

Many of the rocks found in this part of the contact zone are similar to those found in the Glendrian Bay area, but there are some important differences, both in the rock types present and in their age relationships. One of the most important of these is that a number of elongate felsite and microgranodiorite intrusions have been emplaced, more or less up the contact, after the end of high - temperature ( M1 ) metamorphism in the country rocks and during or soon after solidification of the contact zone rocks of the MBG in this area. These M2 intrusions have partly obscured the original M1 contact which corresponded to the wall of the MBG magma chamber.

M1 microgranodiorite directly analogous to that found in the Glendrian Bay area is rather rare. However, microgabbros, dolerites and hybrid rocks have a pillowed and veined contact against microgranodiorite in a small area between 45017026 and 44967026. This contact is notable for the development of microgranodiorite veins which are relatively sharp where they cut the pillowed margins of the more basic and strongly chilled MBG rocks, but rapidly coarsen and become less distinct as they pass inwards and grade into mottled hybrid rocks behind the narrow zone of chilled basic rock at the contact ( Plate 3.20 ). The formation of these structures is discussed further in Chapter 4: for

**Plate 3.20.** M1 microgranodiorite veins cutting chilled contact between MBG microgabbro/dolerite and microgranodiorite, south of Duin Bhain. The veins grade into hybrid rocks further into the intrusion. Field location 42D, grid reference 44997026.

**Plate 3.21.** Quartzite xenolith swarm in hybrid rocks, Field location 20, grid reference 44587031. The largest clasts are about 20cm across.



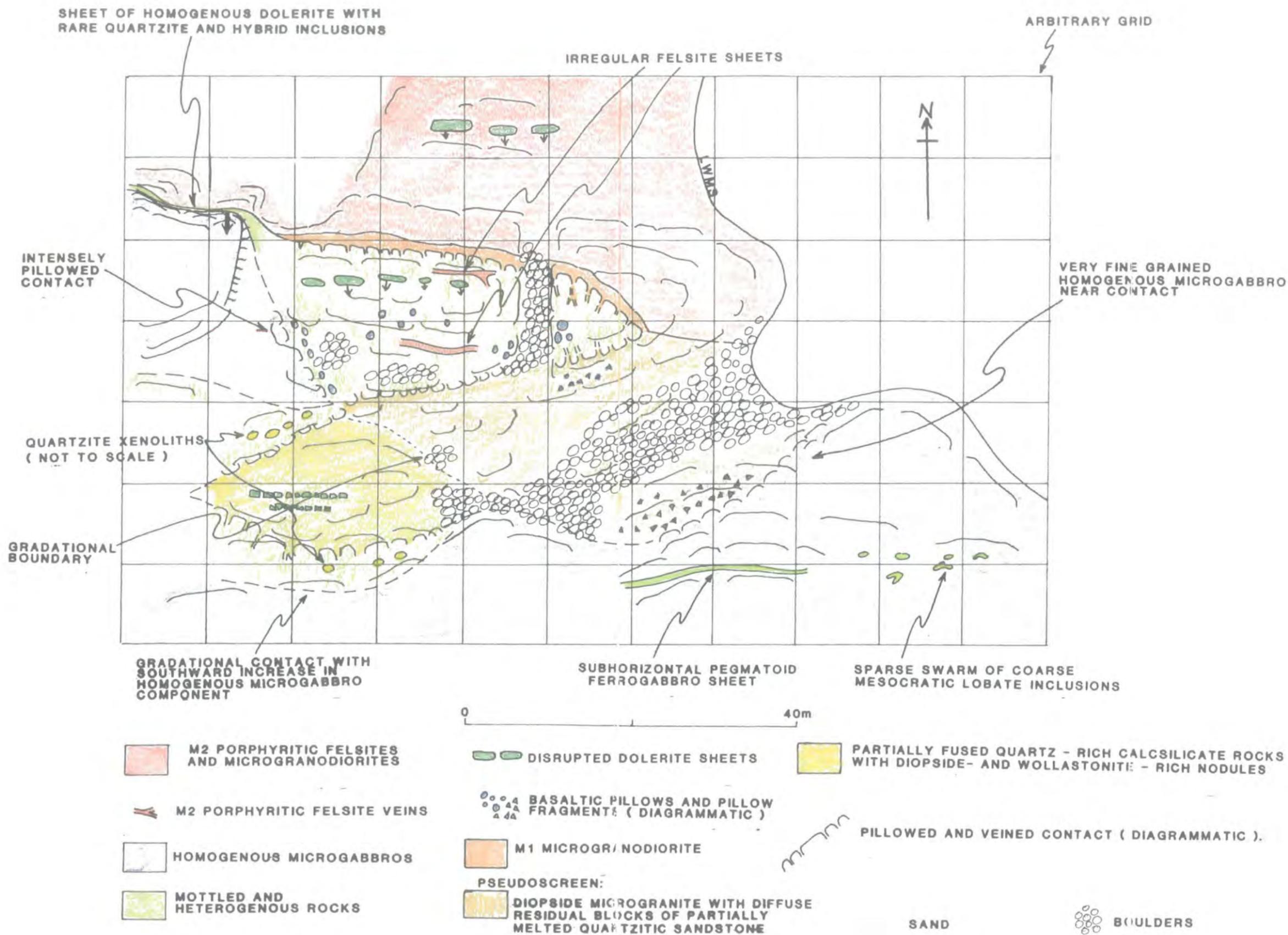
present purposes, they are significant in that they indicate that M1 microgranodiorites are present in this area. Within a metre or so of the contact with the more mafic rocks, the microgranodiorite is intruded by a rather coarse - grained, microgranophyric member of the M2 felsite suite. Further north on the intertidal rock platform south of Duin Bhain, between northings 7070 and 7073, are outcrops of a rather heterogenous granophyric - textured, sparsely porphyritic microgranodiorite with diffuse mafic inclusions. This may represent a screen of M1 microgranodiorite between M2 felsite intrusions.

To the west of Duin Bhain, the heterogenous rocks of the MBG are in direct contact with the country rocks or are only separated from them by M2 felsite intrusions. The country rock - MBG contact, where observed, is sharp. As in the Glendrian Bay area, the character of the rocks close to the contact varies widely. Homogenous fine - grained gabbros occur adjacent to the felsites in the poorly exposed area between 44917027 and 44647031, but elsewhere a zone of heterogenous rocks up to 100m wide is present. These include examples of two rock units not found in the Glendrian Bay area but which are common in other areas of the contact: xenolith swarms and partly fused pseudoscreens of country rocks within the MBG.

The best - developed xenolith swarm in this area is found around 44587031, and is composed of rounded blocks of partially fused diopside - bearing metaquartzite, which occur in a heterogenous, coarse - grained mesocratic host ( Plate 3.21 ). This is coarser grained and more leucocratic than the surrounding MBG hybrid rocks. Within the swarm, which is several metres across, as much as half of the rock is composed of xenoliths, which are all of the same lithology. A more widely dispersed xenolith swarm occurs in homogenous MBG microgabbros in 451701, where lobate inclusions of coarse mesocratic rock have granular quartzite hornfels cores.

Only one pseudoscreen of country rocks is present within the MBG in this sector of the contact zone. This is a body some 30m wide and at least 100m long ( its eastern end is submerged ) on the rock platform south of Duin Bhain. A detailed map of this area is shown in Fig. 3.6. This body resembles a true screen, as defined in Bailey et al. ( 1924 ), insofar as it separates igneous rocks which are chilled against it. This chilling takes the form of pillowed chilled margins which are extensively disrupted by microgranitoid veins. These latter penetrate the chilled margin and grade into the microgabbros as increasingly diffuse areas of mottled hybrid rock, in a manner similar to the

FIG. 3.6. DETAIL MAP OF THE CONTACT ZONE SOUTHEAST OF DUIN BHAIN, SHOWING FIELD RELATIONSHIPS OF THE PSEUDOSCREEN WITHIN THE MARGINAL BORDER GROUP.



veins at the microgranodiorite contact described above ( Plate 3.20 ). The more homogenous, more completely fused parts of the screen also contain small, very strongly chilled basaltic pillows and pillow fragments. However, the mottled heterogenous hybrid rocks around the pseudoscreen form a single rock unit and there is no evidence to suggest that they form more than one intrusion. It appears that the partly to completely fused rocks of this pseudoscreen were displaced into the magmas of the MBG which initially chilled against it but subsequently heated and largely melted it, before both the pseudoscreen and the surrounding magmas were quenched at the end of M1. This quenching is well developed in the granitoid anatectic rocks in the screen: these are variolitic - textured rocks with residual quartz and a quench crystal assemblage which includes quartz pseudomorphs after tridymite, diopside and opaques, as well as an earlier generation of magmatic diopside. The mineralogy of these rocks is discussed further in Chapter 4. The remainder of the pseudoscreen shows lower degrees of partial melting and is a nodular calc - silicate rock with diopside- and wollastonite - rich nodules ( commonly zoned ) in a quartz - rich host. Interstitial partial melts were present in all of these lithologies and were rapidly cooled at the end of M1 to diopside and/or wollastonite and tridymite ( pseudomorphed by quartz ) crystals set in a devitrified quartzofeldspathic interstitial glass. Two early ( pre - or syn - M1 ? ) basaltic sheets cutting these rocks have been intensely veined by the host but have remained essentially intact, indicating that there was no large - scale melt movement in this part of the screen ( Plate 3.22 ).

Within the limits of the exposure, both margins of the pseudoscreen appear to be either steeply inclined or subvertical. The orientation of the boundary between the diopside microgranite and the more refractory calc - silicate rocks is not clear.

The other MBG rocks in this area are broadly similar to those outcropping around Glendrian Bay, except for the following points:

- 1). Coarse grained, relatively leucocratic rocks commonly show fairly well - developed, although recrystallised, pillow structures, particularly along the western part of the exposed contact beyond 44647031. Plate 3.23 shows a microdioritic pillow in a granodioritic host, with a mottled core apparently produced by mixing between dioritic magma and granodioritic magma injected into the pillow through fractures in the chilled outer rind of the pillow, in a manner reminiscent of the larger - scale veins injected into the MBG microgabbros and hybrids at the eastern end of this area (

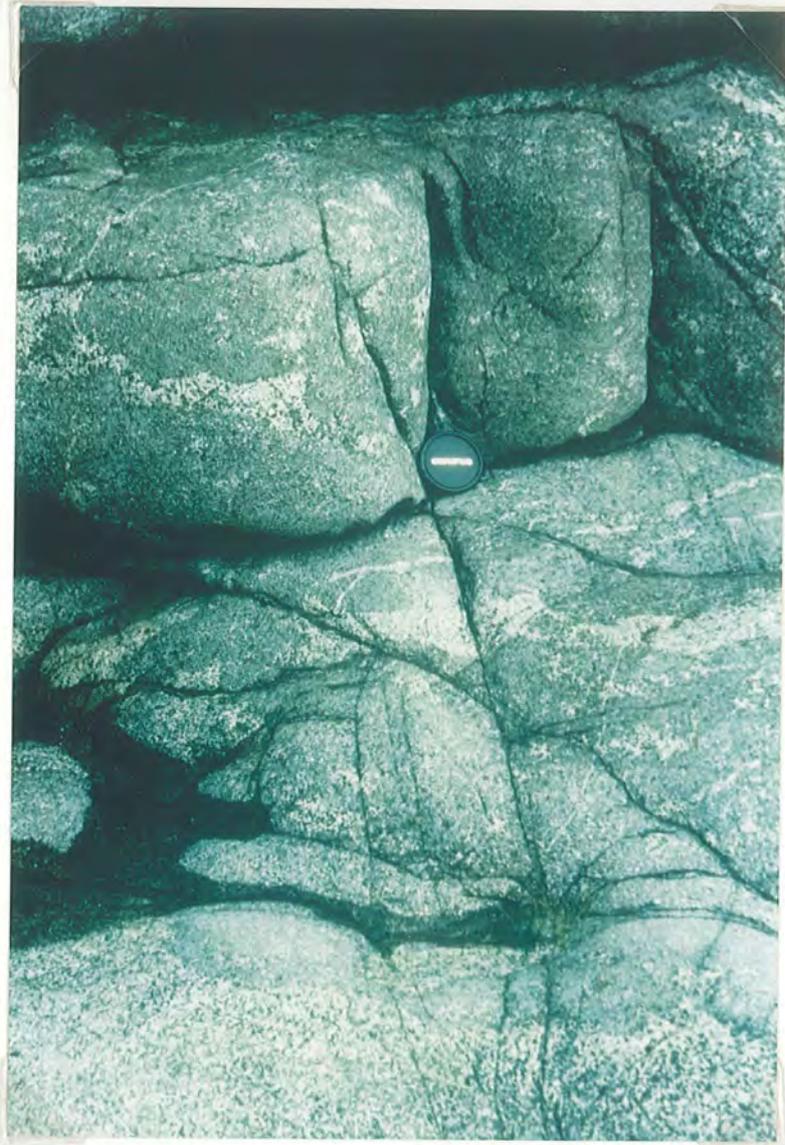


**Plate 3.22.** Brecciated dolerite sheet in partially fused metasediments within pseudoscreen south of Duin Bhain, field location 42, grid reference 44977024. The dolerite sheet predates the melting of the pseudoscreen following its formation by fault movements ( see text ) and has been brecciated by melts not related to its own emplacement ( it is not, therefore, a true backveined sheet ). The fact that the original geometry of the sheet is still recognisable implies that the scale of melt movement in this part of the pseudoscreen was small.



**Plate 3.23.** Microdioritic pillows in granodioritic host rock, forming part of the heterogenous hybrid outcrops around 44627031. The chilled margin of the pillow in the foreground cracked whilst the interior was still molten, admitting diffuse granodioritic veins into the interior of the pillow.

**Plate 3.24.** Irregular lobate pillow, with crenulate chilled margins, in faintly mottled heterogenous hybrid rocks. The pillow itself is composed of homogenous basic rock. Adjacent to pseudoscreen south of Duin Bhain, grid reference 44977025.



compare Plate 3.20 ).

2). Evidence for multiple stages of mixing is again common, in the shape of basaltic or doleritic pillows in mottled hybrid rocks ( Plate 3.24 ) and lobate inclusions of coarse heterogenous mesocratic rock in the microgabbros. However, many of the basic pillows are porphyritic, unlike the aphyric rocks in the Glendrian Bay area. Typically, less than 1% to as much as 5% of these rocks is composed of short, zoned labradoritic feldspar phenocrysts, some of which have rounded or corroded cores. This is also a feature of many of the cone sheets in the area.

3). In addition to coarse mottled hybrid and ferrogabbro veins, the MBG rocks in this area are cut by medium to coarse grained irregular granodiorite veins and larger subhorizontal to slightly north - dipping sheets. The largest example is in grid square 441703 where a sheet 70m long and 2 - 3m thick is the centre of a network of smaller horizontal sheet intrusions and abundant net - veining of the host mottled rocks. The irregular geometry, unchilled margins and coarse grain size of these rocks suggests that they were emplaced when the host MBG rocks were largely or entirely solidified but still hot. The main sheet in 441703 grades into more coarsely and strongly mottled rocks to the west, suggesting that the host may still have been partially molten at the time of emplacement. In contrast, veins of similar composition in MBG rocks in the east of the area, south of Duin Bhain, are porphyritic felsites and post - quench ( i.e M2 ) in age.

4). The quenching event which separates M1 and M2 is not as well developed in the western part of this area as it is around Duin Bhain or in the Glendrian Bay sector. Pelitic hornfels around 44267045 have a well - preserved fine - grained M1 assemblage, suggesting rapid cooling, and interstitial microgranophyre is common in the MBG rocks themselves, but devitrified interstitial glasses are absent. As noted in Section 3.2.2, the cooling of the MBG appears to have been a much more protracted process than that of the M1 contact aureole: these rocks could simply be a more extreme case of this, or have been subjected to a longer period of post - quench recrystallisation, or recrystallisation at higher temperatures ( Lofgren 1971 ). A boundary between more rapidly and more slowly cooled rocks may be present in the northernmost outcrops in grid square 443704, where microgranodiorite with a pillowed and veined contact against mottled hybrid rocks grades north into a finer - grained porphyritic microgranodiorite. This would imply that rapid cooling occurred up to the contact zone but proceeded more slowly in the microgranodiorite and other rocks to the south.

Additional evidence for more intense post - M1 metamorphism in the rocks at the western end of this sector comes from a sample described in Day ( 1985 ). This is a hybrid rock from grid square 441704, in which secondary chlorite, replacing primary clinopyroxene, is itself overprinted by the growth of clusters of bladed hornblende crystals. This suggests slight prograde metamorphism after the end of M1, either during M2 or later. Prograde metamorphism during M2 also occurs in other areas of the contact ( see summary in section 3.5 ): possible causes of this are discussed in Chapter 5.

### **3.2.3.2. Country rocks adjacent to the Hypersthene Gabbro between Sanna Point and Duin Bhain.**

The geology of the pre - Tertiary country rocks in this sector of the contact is almost entirely obscured by cone sheets, but two separate groups of rocks can be distinguished. Firstly, to the east of a group of prominent NW - SE and NNW - SSE gullies between eastings 443 and 445 ( see Map 3 ), screens of high - grade pelitic, calc - silicate and calc - quartzite hornfels are present between cone sheets and define three domains:

- 1). Close to the contact, in grid squares 445703 and 446703 are a variety of calc - silicate ( wollastonite - bearing but strongly altered during M2 to grossular and idocrase rich rock ) and diopside quartzite rocks: the latter are similar to the xenoliths in the adjacent xenolith swarm in the MBG. Bedding in these rocks at 44567032 dips away from the contact at  $25^{\circ}$ . These rocks may also be correlated, on grounds of similar composition, with rocks in the pseudoscreen within the MBG to the east ( Fig 3.6 and Map 3 ).
- 2). North of the above, in a discontinuous belt from easting 445 to the eastern limit of outcrop, are dark grey fine - grained quartzose semi - pelitic hornfels with nodular bands of greenish fine - grained hornfels that dip  $10 - 15^{\circ}$  north. These rocks can be correlated with the nodular, slightly calcareous hornfels at the top of the Pabba Beds in the Glendrian Bay area ( Chapter 2 ). Small disrupted screens of similar rocks occur between the porphyritic felsite intrusions to the south.
- 3). To the north again, in an area that is almost entirely occupied by cone sheets and later porphyritic felsite sheets, are isolated screens of pale green diopside quartzite hornfels.

The orientation of the bedding within these screens suggests that these rocks have an overall outward dip, away from the Hypersthene Gabbro. Units (2) and (3) above can be correlated with the upper part of the Pabba Beds and the Scalpa sandstone in the Glendrian Bay area, but the position of Unit (3) is anomalous. Apart from the calcsilicate rock at 44537032 this unit also resembles the Scalpa sandstone and the rocks in the pseudoscreen, but apparently dips beneath unit (2). One possible interpretation is that a fault, downthrowing to the south was present, running parallel to the present M2 felsite - MBG contact. There is however no direct evidence for the presence of such a fault, perhaps because of subsequent intrusion of cone sheets and felsites.

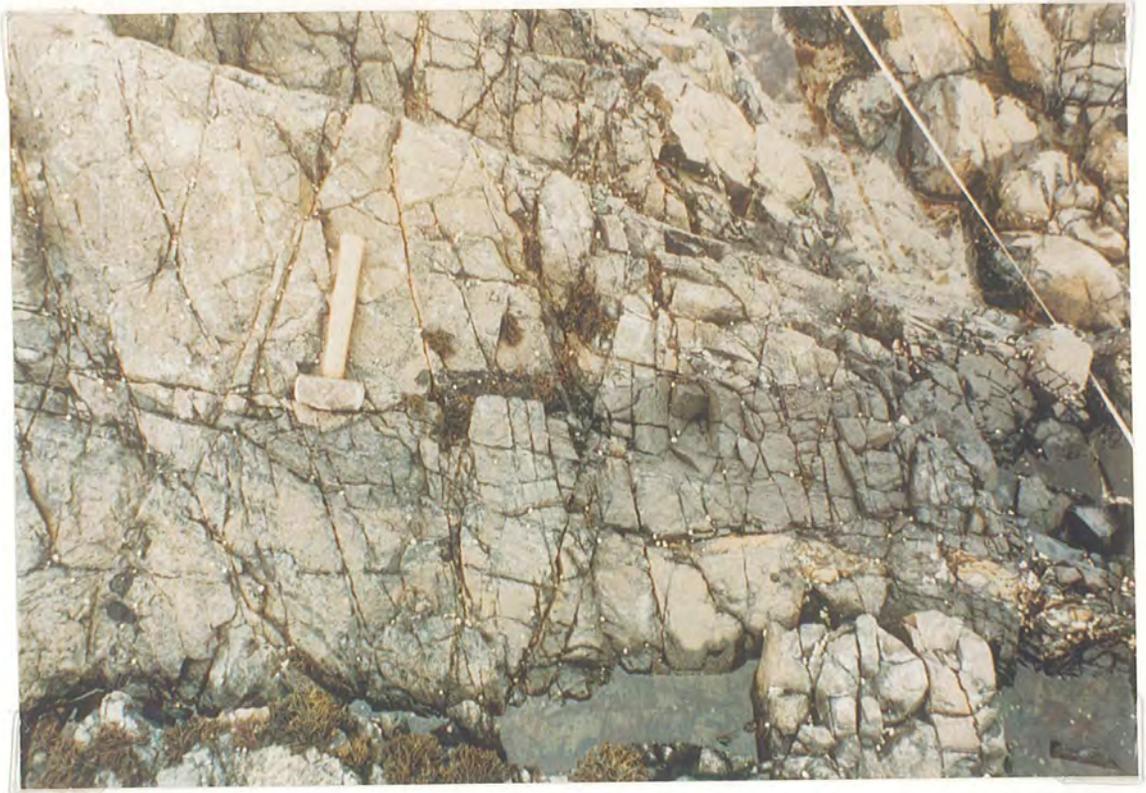
The second group of sediments is represented solely by an isolated outcrop some 30m wide, right on the coast at about 44267045. This is composed of banded, very dark pelites, similar to the FeAl - rich laminated pelites of the Middle Pabba Beds. Similar dark pelites occur as xenoliths in the M2 felsites further east.

### **3.2.3.3. Cone Sheets and Felsite Intrusions between Sanna Point and Duin Bhain.**

There is better evidence for the relative ages of the cone sheets in this area, as compared to that on the far side of Glendrian Bay. Many are strongly hornfelsed at outcrop and have a granular texture, and around 44517034 the cone sheets are truncated by the MBG, implying an M1 or earlier age. Elsewhere, however, porphyritic dolerite hornfels sheets cut the MBG ( for example at 44247045 and 44977026 ), although these are still somewhat hornfelsed and backveined in some cases as well, implying an M2 age. The area of homogenous microgabbro between the felsites and the pseudoscreen, at the western end of Fig. 3.6, appears to have south - dipping upper and lower margins ( the former is present in cliffs around 44917024 ), and may represent a cone sheet - like intrusion emplaced into partially molten MBG rocks. Most of the cone sheets are cut by the large M2 felsite intrusions, or sheets and veins extending out from these, but a number of south - dipping and subvertical dolerite sheets cut the felsites, particularly at the northern edge and western limit of the main felsite intrusions ( Map 3 ). A number of less regular, north - dipping doleritic and basaltic sheets and veins also cut the felsites but are absent further north, where all the cone sheets dip south. These anomalous intrusions are commonly associated with shear zones cutting the felsites ( see below ): an example is shown in Plate 3.25. A few planar, vertical east - west trending composite and hybrid acid - basic dykes also cut the felsites around 450703. These are very unlike the exclusively basic

**Plate 3.25.** Irregular north - dipping dolerite sheet in M2 felsite intrusion south of Duin Bhain, grid reference 44987028 approx. The margins of the sheet are sheared in sub - parallel fault zones. Indistinct north - dipping shear zones also occur in the host rocks.

**Plate 3.26.** Hypidiomorphic granular textured plagioclase - augite - magnetite ( microdioritic ) glomerocryst in M2 felsite, sample 42F2. Probably of xenolithic or residual origin ( see Chapter 4; the same glomerocryst is show in plane - polarised light in Plate 4.1 ). Crossed polars, field of view 3mm.



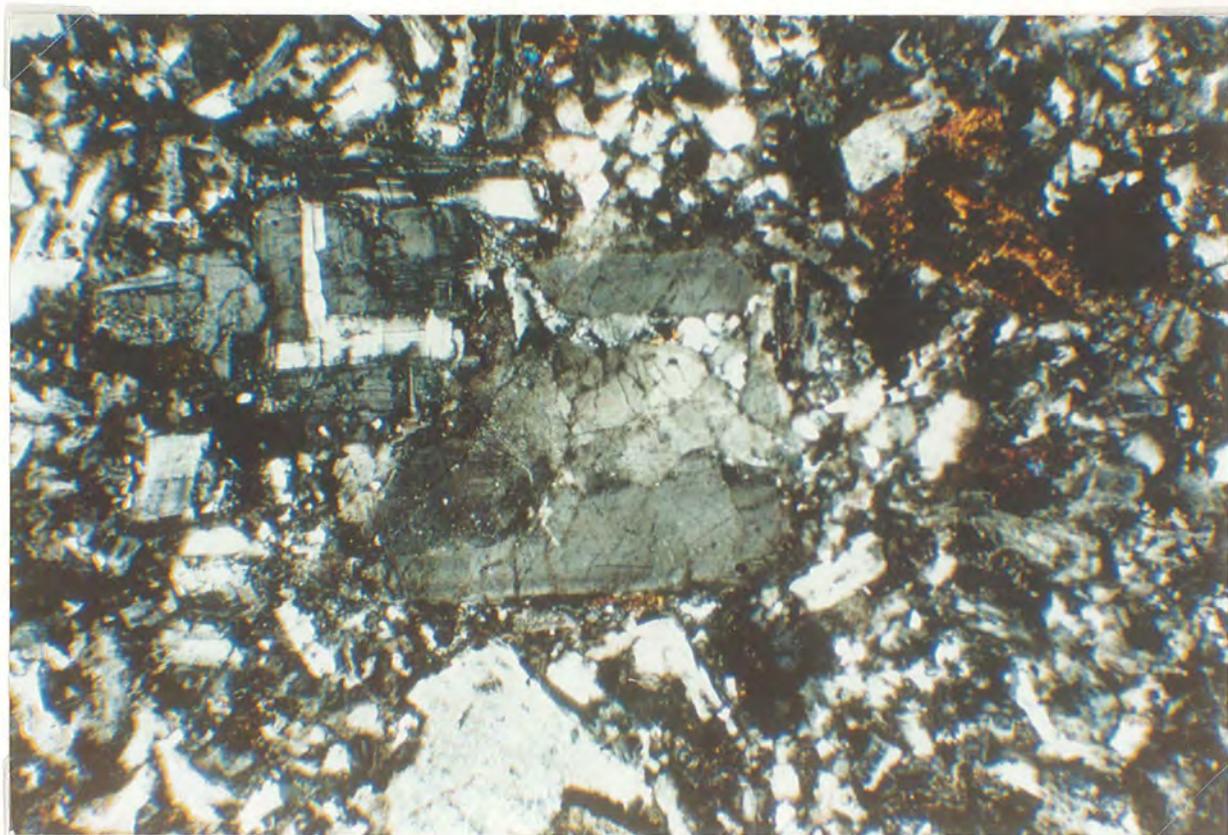
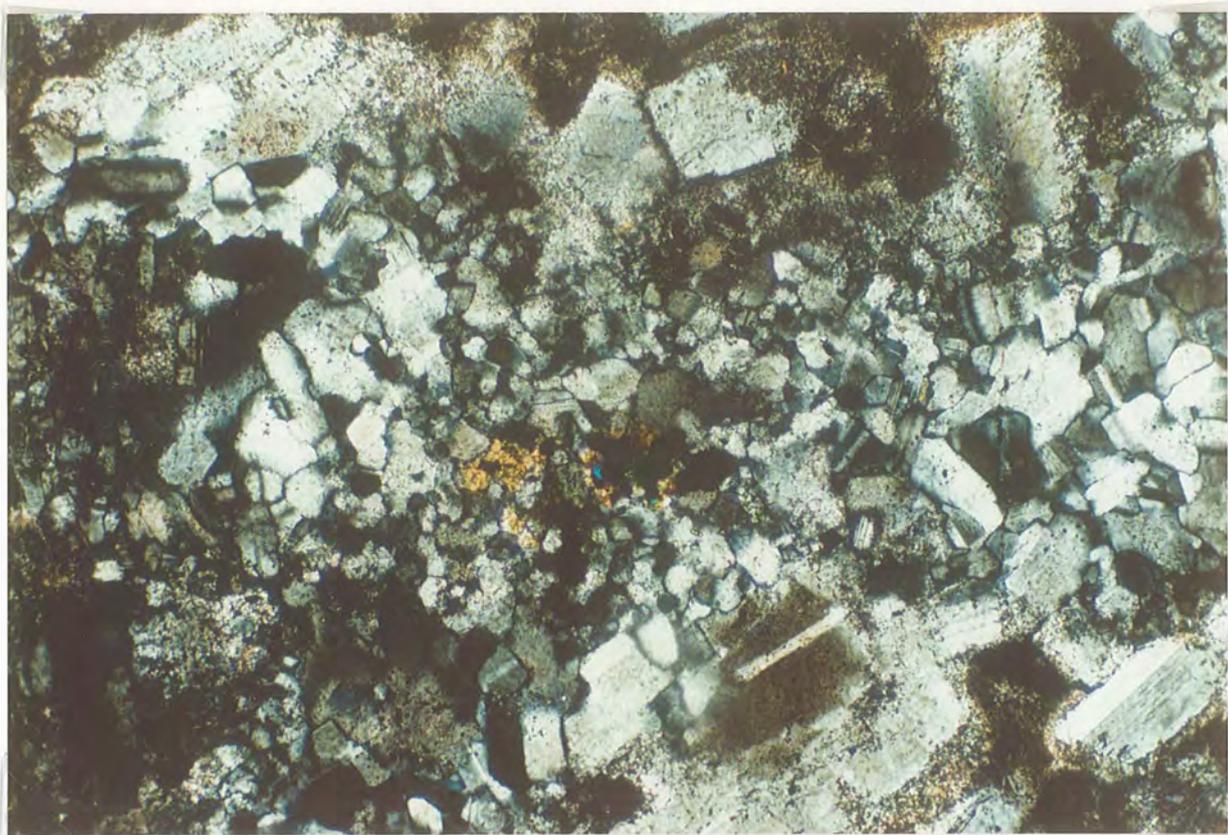
cone sheet intrusions and may be more plausibly related to the Sgurr nam Meann hybrid intrusion. This would be consistent with their fresh, unaltered appearance.

Multiple intrusions of porphyritic felsite and microgranodiorite emplaced during M2 ( from cross - cutting relationships ) occupy a tract 500m long and up to 100m wide in the eastern part of the Sanna - Duin Bhain sector. M2 felsites also occur widely as veins and south - dipping inclined sheets ( cone sheets ? ) in the country rocks to the north and west and, less frequently, in the MBG rocks as well. The eastern limit of the main tract is formed by the sea, and the intrusions probably extend east under the sea for at least a few hundred metres more. Overall, the tract has steep contacts against the older rocks to north and south but the two margins do not match up, implying that emplacement of the felsites was not purely dilatational. Doleritic and metasedimentary xenoliths are common: although some of these can be shown to be the remnants of disrupted screens, from their occurrence in elongate trails at the margins of individual felsite sheets, their presence implies that stoping and possibly brecciation of the margins took place during felsite intrusion. As noted above, the principal rock type missing from the M1 MBG contact zone is microgranodiorite and it is possible that its absence may be explained by erosion or structural cutting - out at the present level of exposure during emplacement of the M2 felsites.

A large number of M2 felsitic and slightly coarser microgranodioritic intrusions are present. The largest form sub - parallel vertical sheets within the tract mentioned above, but the smaller intrusions resemble cone sheets or are irregular ( see above ). In thin section the rocks are characterised by a well - developed porphyritic texture and, variously, microgranophyric, variolitic or spherulitic devitrified groundmasses. Although euhedral phenocrysts and quench crystals of feldspar, augite and opaques occur in the microgranophyric rocks and in some others, the bulk of the porphyrocrysts show varying degrees of corrosion and resorption. The porphyrocryst assemblage is composed of plagioclase and variable but much smaller amounts of augite and opaques. Plagioclase - augite and other glomerocrysts are commonly present. Many of these have sub - granular hypidiomorphic textures ( Plate 3.26 ), suggesting that they are microxenoliths rather than magmatic glomerocrysts. The augites in some slides are pseudomorphed by turbid aggregates rich in very fine - grained granular opaques. Pre - magmatic deformation is recorded in a few microxenoliths with granular - textured annealed shear zones ( Plate 3.27A ), while less severe intracrystalline deformation , marked by local granulation, subgrain development and strained extinction is almost ubiquitous in

**Plate 3.27A.** Microxenolith containing granular annealed shear zone, in undeformed M2 felsite, sample 42F2. Crossed polars, field of view 3mm.

**Plate 3.27B.** Large feldspar porphyrocryst with intense internal straining and subgrain development in undeformed M2 felsite, sample 42C. Crossed polars, field of view 3.0mm.



**Plate 3.28.** Tensile fractures forming hydrothermal vein network in devitrified M2 felsite highlighted by alteration haloes around the hydrothermal veins. South of Duin Bhain, grid reference 44957028.



the corroded feldspars ( Plate 3.27B ). This may also be a relict texture because the deformation is not seen in the groundmass. A variety of sieve and negative sieve textures are visible in the feldspar porphyrocrysts ( see section 3.2.6 and Chapter 4 for descriptions of these textures ), with distinct later quench overgrowths. **The textural evidence therefore suggests that the bulk of the larger porphyrocrysts and glomerocrysts in these rocks are relict and inherited from an incomplete melting episode just prior to emplacement of the M2 felsite and allied intrusions.** This is not to say that all of the material in these rocks is of anatectic origin: the variety of feldspar porphyrocryst textures in particular rocks suggests that magma - mixing has taken place. The source of these rocks and their evolution is discussed further in Chapter 4. For the present, although by definition melting of pre - existing wall rocks at any one level of exposure ended at the M1/M2 transition, **at that level of exposure**, it appears that melting in the source region of the felsitic magmas, at depth, continued into the period of M2 metamorphism at the present level of exposure. This suggests that the M1 to M2 transition may have been diachronous vertically, but on a greater scale than presently exposed. Although they are M2 in age, alteration in most of these felsic intrusions is relatively slight and mainly involves the augite porphyrocrysts and varying degrees of devitrification and recrystallisation of the quenched interstitial material . Like M2 metamorphism elsewhere ( see especially section 3.2.5.5 ), alteration in these rocks is concentrated along fractures which are made very conspicuous in spherulitic - textured felsites around 44957028 by bleaching associated with more advanced devitrification around the fractures ( Plate 3.28 ).

#### **3.2.3.4. Deformation in the Sanna Point - Duin Bhain area and a summary of the overall age relationships.**

Deformation in this area is best developed towards the eastern end of the exposed contact. As is the case around the eastern side of Glendrian Bay, any early deformation is largely obscured by the later cone sheets. However the Jurassic sediments do appear to have an overall northerly dip corresponding to the outward dips in the Glendrian Bay area, where faults associated with the outward tilting of the Jurassic sequence were active at an early stage in the history of the area ( Fig. 3.5 ). If the correlation of the grey and green pelitic rocks with the Upper Pabba Beds, and that of the calc - silicates and diopside quartzites with the Scalpa Sandstone is valid ( see above and Chapter 4 ), downfaulting of rocks towards the contact may have occurred during M1 in a manner to that seen much more clearly on the southern margin of the intrusion ( section 3.2.5.1 ).

M2 deformation is similar to that in the Glendrian Bay area, with undulating shear zones striking parallel to the contact overall and dipping north or, less commonly, south at angles of 30° or so. The zones of most intense deformation are up to 10cm thick and composed of dark, sometimes banded, fine - grained to flinty annealed cataclastic rock. The smaller shear zones in particular have an irregular, wispy appearance ( Plate 3.29 ). The host to these distinct shear zones is often formed by mylonitic to weakly sheared microgranodiorite or felsite which forms more diffuse regions of deformed rock ( see Chapter 7 ). Evidence to be discussed in Chapter 7 suggests that these shear zones are mainly thrusts. As noted above, doleritic sheets and veins are associated with the thrusts ( Plate 3.25, above ): some of these have deformed and brecciated chilled margins indicative of syn - thrust emplacement ( Plate 3.30 ). Post - deformational hydrothermal veins and recrystallisation of the crush - rocks imply that the deformation took place during M2.

The latest element of deformation in the area is a group of north - south and NNW - SSE trending subvertical faults. These may belong to the same set of structures as similar faults of the same age on the southern margin of the intrusion ( section 3.2.5.4 ). A number of dykes in the area with similar orientation also have very late relative ages: they cut all other rock types ( see Map 3 ) and show very little alteration.

The age relationships of the rocks in the Duin Bhain - Sanna Point area are summarised in Fig. 3.7.

### **3.2.4. Ardnamurchan Point To An Acairseid.**

#### **3.2.4.1. The geometry of the western margin of the MBG.**

The geometry of the outer contact of the MBG and of the Hypersthene Gabbro as a whole has been greatly complicated in this area by later movements on a series of NNW - SSE trending faults which cut all components of the pluton ( see Map 1 and Fig.3.29 ). The most important of these in the context of the MBG is the fault running from Port Min ( Grid Ref. 417661 ) to Port Choinnich ( 423643 ). This appears to be a normal fault, downthrowing to the east, which has in effect repeated the steeply west - dipping outer contact of the intrusion. The latter is actually only exposed in the hanging -wall block, on Druim na Cloise, but also appears to lie only just offshore from the western coast of Ardnamurchan, in the footwall of the fault. The evidence for its presence offshore is the

**Plate 3.29.** Indistinct north - dipping shear zones ( wispy pale grey bands underneath irregular dolerite sheet below hammer, and at top of field of view ) cutting M2 felsite south of Duin Bhain. Grid reference 44957028.

**Plate 3.30.** Deformed chilled margin of dolerite sheet in M2 felsite, south of Duin Bhain ( sample 43L3 ). The presence of fragments of very fine grained chilled rock and of sheared wall - rock felsite in the less - deformed chilled margin indicates that the deformation was synchronous with the emplacement of the dolerite. Plane polarised light, field of view 3mm.

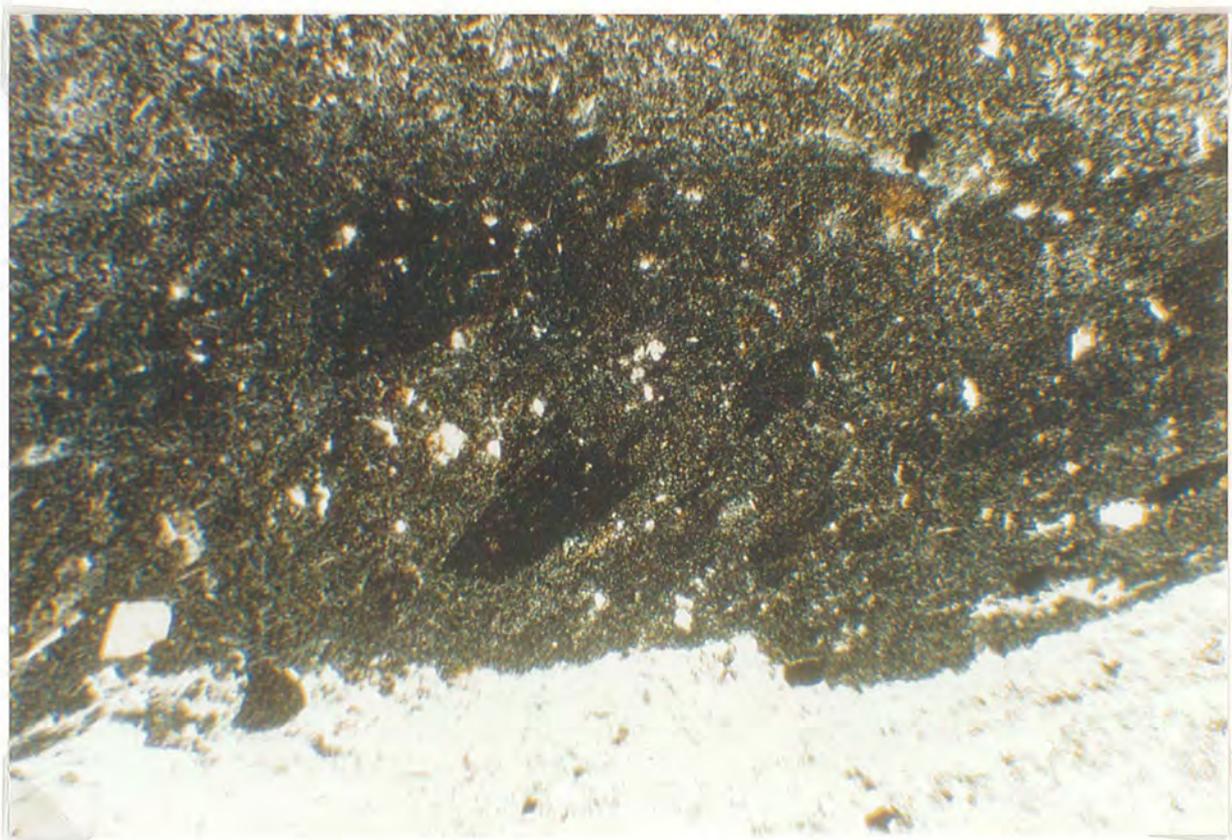
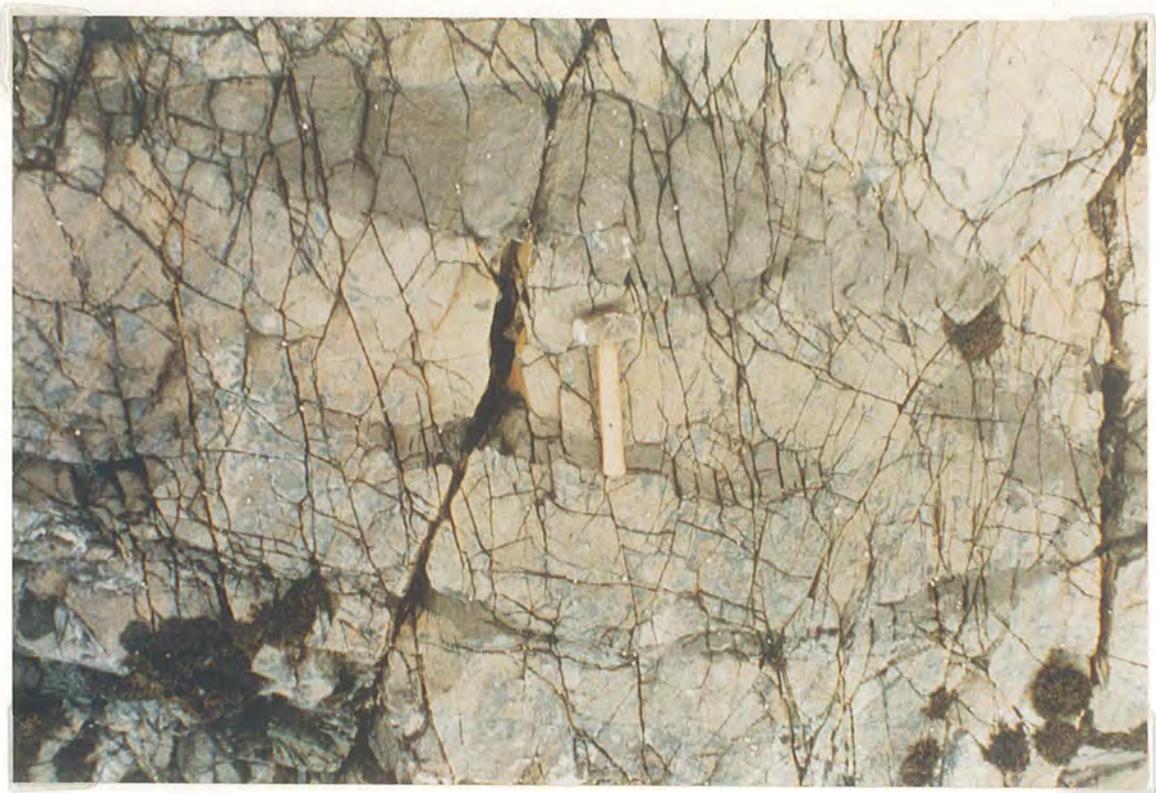
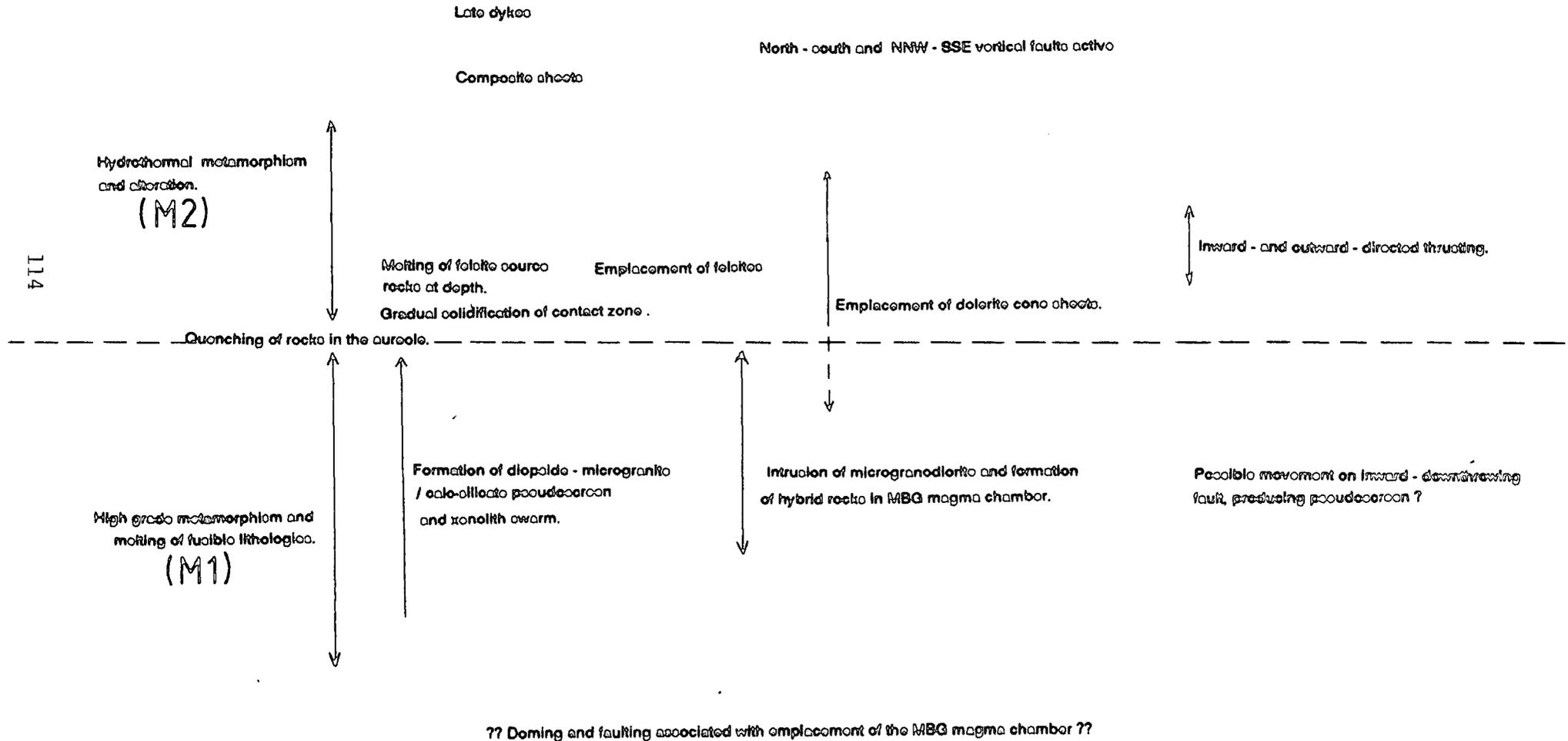


Fig. 3.7. Relative ages of intrusions, metamorphism and deformation in the Duin Bhain - Sanna sector.



development along this coast of a suite of rocks comparable to that found in the contact zone of the MBG elsewhere around the pluton. The suite, of xenolith swarms, mottled hybrid rocks and coarse to pegmatoid - textured ferrogabbroic to ferrodioritic sheets, veins and lobate inclusions, is developed in three parts of this sector of the contact, at Ardnamurchan Point ( which actually lies to the east of the projection northwards of the fault, but the displacement on this either dies out to the north or is transferred to faults east of the Point along an east - west shear zone through Briaghlan: see Map 1 ); on Chorrachadh Mhor ( Grid Ref. 414662 ); and along the western side of Gharblach Mhor between 415661 and 418651. In all three, there is a zone 30 to 100 metres wide dominated by contact zone rocks, which passes inwards into homogenous microgabbroites and dolerites which are the dominant MBG rock type in this as in other areas. The eastern limit of the contact zone, and hence the contact itself, appears to be steeply - dipping, notably in the cliffs at the northern end of Gharblach Mhor.

Similar cliff exposures in the 100m high cliffs on the north - western side of An Acairseid, around 431637, show the contact on the eastern side of Druim na Cloise to dip very steeply southwestwards, at at least 70°. Wells ( 1954 ) considered that the contact in this area dipped south - west at a much shallower angle, on the basis of supposed occurrences of MBG rocks between the Port Min - Port Choinnich fault and the western side of the country rocks on Druim na Cloise. However, except at the northern extremity of Druim na Cloise, no outcrops of convincing MBG lithologies were found here during the course of this study. Fine grained basic hornfels were found at the foot of the cliffs on the western side of Druim na Cloise, around 423645, but their intensely hornfelsed, homogenous character suggests that they are more likely to be isolated outcrops of cone sheets cutting the country rocks, although no actual contacts were found. It therefore seems that the contact does indeed dip steeply outwards overall. The country rocks at the northern end of Druim na Cloise ( Grid Ref. 419654 ) are extensively melted and disrupted, in contrast to those further south, which dip uniformly south - west, away from the contact, and only show evidence for extensive melting in the more fusible lithologies. The rocks at the northern end of Druim na Cloise may therefore form a pseudoscreen partially enclosed by the MBG, whilst the contact to the south appears to represent the main outer contact of the pluton.

#### **3.2.4.2. Contact Zone Rocks along the western margin of the MBG.**

The xenolith swarms along the western coast are commonly smaller and less concentrated than many elsewhere along the contact zone of the MBG, but like them are dominated by particular lithologies. The largest swarm, on the west side of Gharblach Mhor ( Grid Ref. 416655 ) is composed of diopside - bearing quartzite hornfelses with interstitial patches and veins of intergrown quartz and feldspar which correspond to crystallised partial melts. However, most of the xenoliths on Gharblach Mhor are hydrothermally altered hypersthene - cordierite - plagioclase - quartz hornfelses with interstitial microgranophyre, corresponding to refractory quartz- and Fe - rich lithologies from the ferruginous siltstones of the Lower Pabba Beds ( Chapter 2 ). At Ardnamurchan Point, quartz - rich xenoliths with interstitial microgranophyre, possibly analogous to the residual quartzites in the rheomorphic breccias at Glendrian Bay, are the dominant xenolith type. All the xenoliths, like the other rock types in the area, have been affected by subsequent hydrous alteration and subsolidus recrystallisation ( Plate 3.31 ). The xenoliths are usually enclosed in homogenous microgabbros or in coarse ferrogabbros or ferrodiorites: only in the case of the large diopside - quartzite swarm at 419655 do large amounts of coarse mottled hybrid granitoid rock enclose the xenoliths. A few granulitic - textured, fine - grained, homogenous olivine - bearing metabasic xenoliths are also present at Ardnamurchan Point.

Microgranodiorite and other fine - grained felsic contact zone rocks are conspicuously absent from the exposures along the western coast of Ardnamurchan, although it is of course possible that they may be present offshore, closer to the contact. The dominant contact zone rock types exposed are quartz microgabbros or microdiorites and a suite of coarse to pegmatoid - textured and opaque - rich ferrogabbros and ferrodiorites. The latter are very similar in the field and petrographically to those found on the northern margin of the intrusion, except that quenched interstitial glasses are absent, their place being taken by coarse granophyre. This suggests slower cooling of these rocks in the later stages of their crystallisation. The iron - rich rocks occur in a zone up to 50m wide which at Ardnamurchan Point is inboard of a zone of homogenous quartz microgabbro ( Fig. 3.8 ): elsewhere, they are the closest rocks to the sea. They form veins, pods and irregular sheets in fine - grained, homogenous aphyric microgabbros and dolerites similar to those found further into the MBG ( Plate 3.32 ) and in places form more than 50% of the outcrop. The microgabbros and dolerites are rather richer in olivine and orthopyroxene than many MBG microgabbros but are correlated with them rather than with the Inner Series because of the lower abundances of these minerals , relative to IS rocks, and the lack of other features diagnostic of the Inner Series ( see section 3.3 ), as well as their

**Plate 3.31.** Partly fused, quenched and intensely re - crystallised feldspathic quartzite xenolith in MBG rocks, Ardnamurchan Point ( grid reference 41576718 ). Note the patch of relict granophyre at right, also the irregular channels, probably formed by feldspars, in the quartz poikiloblast at the lower left of the field of view. Sample 101/3. Crossed polars, field of view 1.5mm.

**Plate 3.32.** Irregular veins of coarse ferrogabbro in doleritic MBG microgabbro, Ardnamurchan Point. Grid reference 41616717.

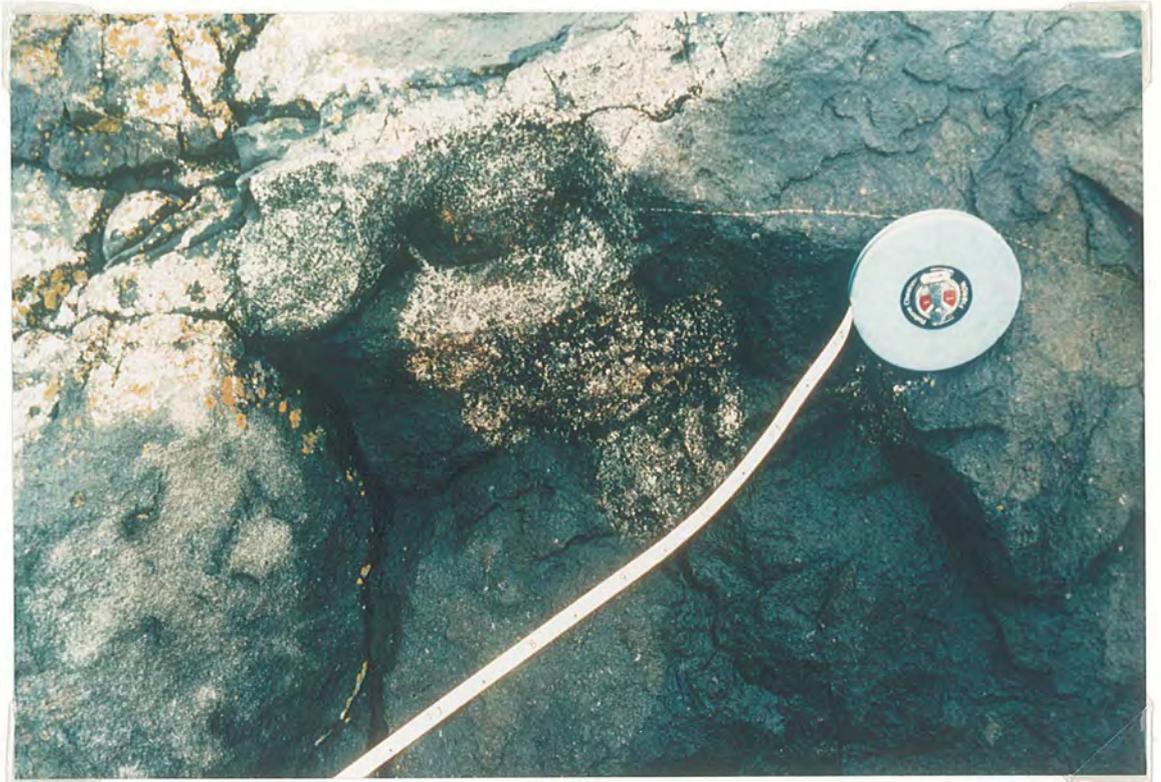
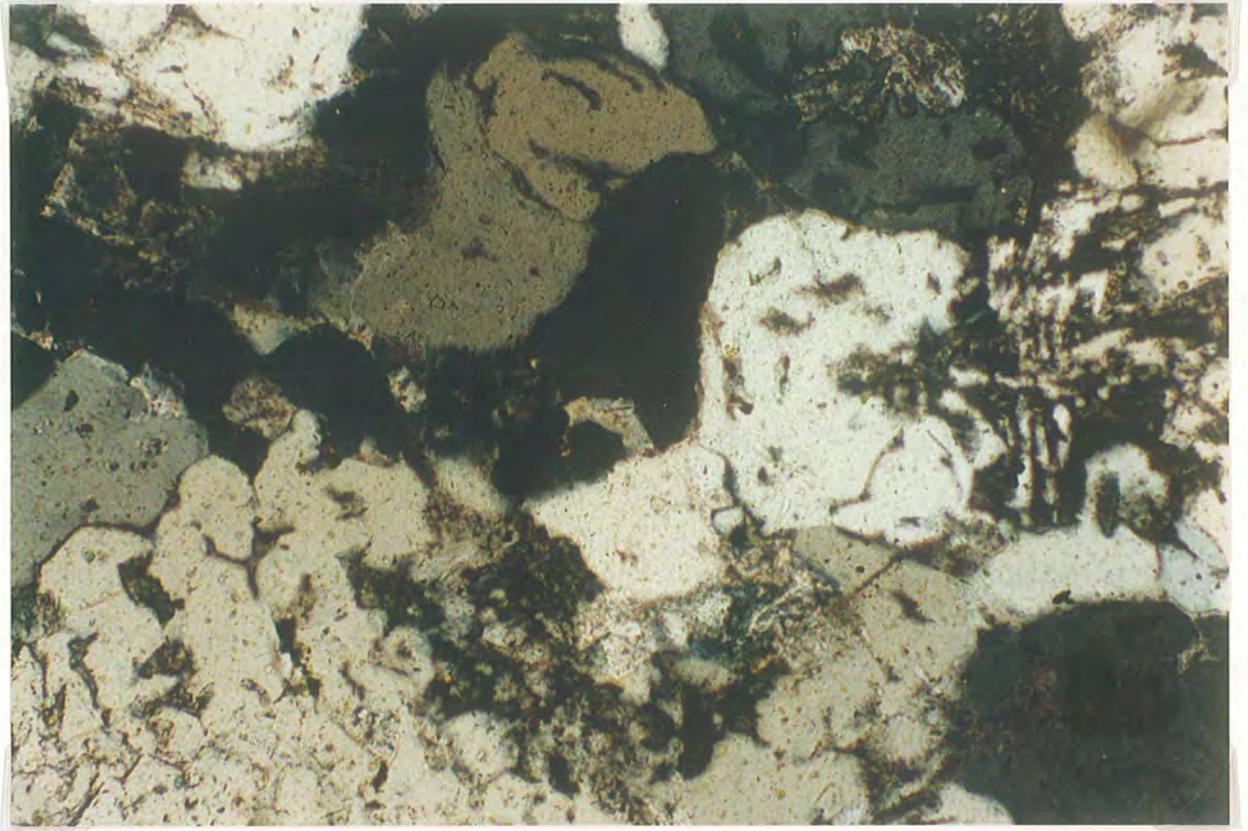


FIG 3.8 SKETCH MAP OF THE CONTACT ZONE OF THE MARGINAL BORDER GROUP AT ARDNAMURCHAN POINT AND THE ADJACENT CONTACT WITH THE INNER SERIES ( SECTION 3.3 ).



age relationships to the ferrogabbros and ferrodiorites. As with similar rocks on the northern margin of the intrusion, the lack of chilled margins and the irregular and lobate forms of the ferrogabbroic and ferrodioritic bodies suggest that they were intruded when their host rocks were still hot and probably partially molten. The highly irregular geometry of some of the veins and pods in particular ( see Plate 3.32 ) suggests that they formed in part by replacement of the host rocks, rather than simply by filling of dilational cracks in the host rocks. The possible origins of this phenomenon are discussed in Chapter 5: for the present, it should be noted that, by analogy with the iron - rich sheets and veins on the northern margin of the intrusion, these ferrogabbros and ferrodiorites should date from the early part of M2, although in the absence of exposures of the contact this cannot be directly confirmed. All the contact zone rocks exposed in this area do however show the effects of hydrothermal metamorphism similar to M2 in character, with the replacement of primary mafic phases by talc + opaques ( olivine ) and chlorite + green amphibole ( pyroxenes ) being especially prevalent.

One rock type is present along the western margin of the intrusion which is virtually absent from the northern contact zone in particular. This is a suite of sparsely plagioclase - phyrlic, olivine - poor dolerite intrusions with unchilled margins against the host contact zone rocks. The smaller examples are sheet - like and dip east at angles of less than 40°: the larger bodies, such as that which forms the south - west tip of Ardnamurchan Point ( Fig. 3.8 ), are not well enough exposed for their geometries to be determined, although they do have east - dipping contacts at top or bottom. Whilst the contacts between these porphyritic dolerites and the microgabbros are planar, the contacts between the dolerites and the ferrogabbros and ferrodiorites are lobate or irregular, and backveining of the dolerites is common. This suggests that these particular inclined sheets were emplaced when the host rocks were still hot and the iron - rich rocks were still partially molten: hence, early on in M2. This is consistent with the intense moderate - temperature, M2 - type hydrous alteration present in these porphyritic dolerites. They therefore seem to be equivalent to the early M2, sparsely porphyritic dolerite cone sheets found to the north - east of the pluton in the Glendrian Bay area. It should be noted that these rocks are petrographically distinct from the much more strongly feldspar - phyrlic inclined sheets which cut Inner Series rocks further inland and also from the coarse, feldspar porphyrocryst - bearing olivine gabbroites which occur to the west of the Port Min - Port Choinnich fault at the southern end of Gharblach Mhor and around Port Choinnich itself. The latter are very similar to the olivine gabbroites which make up the bulk of

the Inner Series; both groups of rocks will be discussed further in sections 3.3 and 3.4.

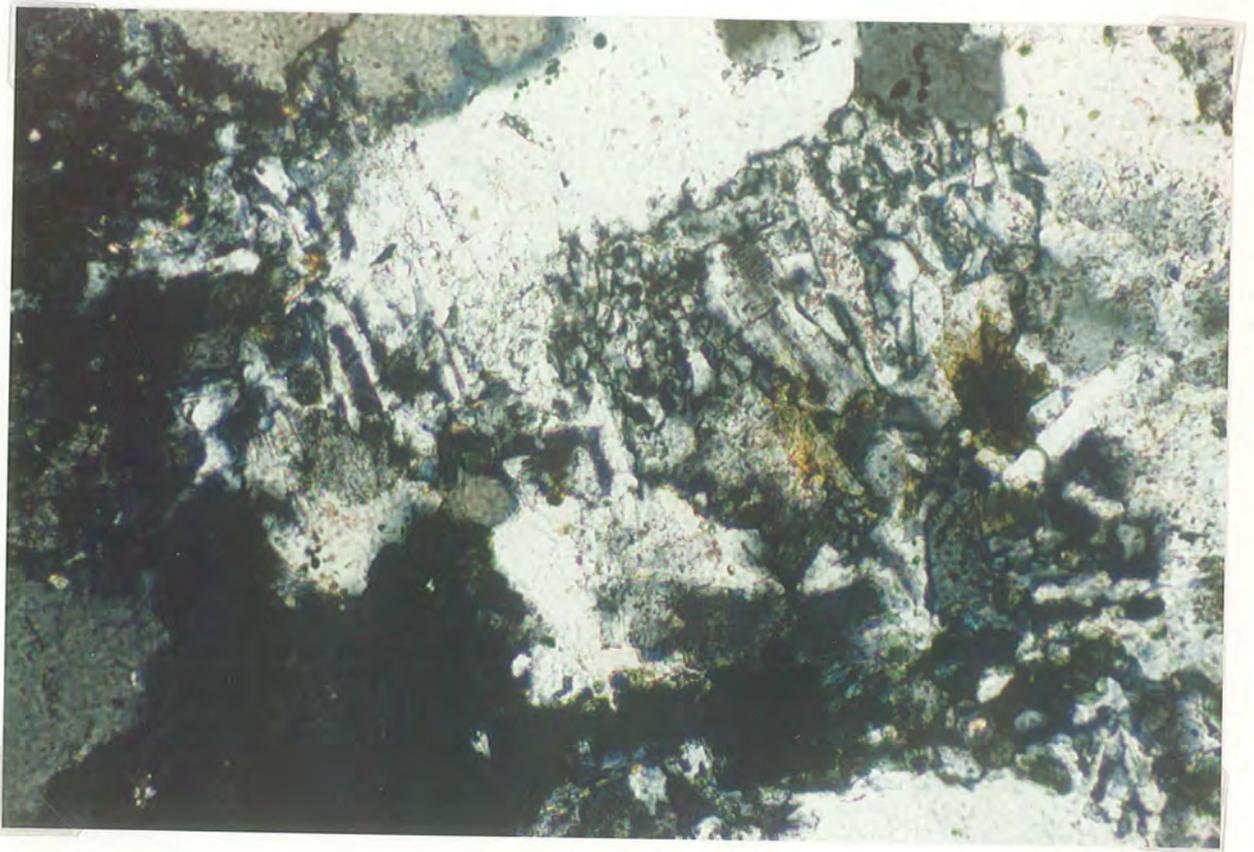
The contact zone on Druim na Cloise is significantly different from that on the west coast of Ardnamurchan, in that the coarse - grained heterogenous rocks typical of the latter are largely absent. Except at the northern end of the country rock outcrop on Druim na Cloise and in the cliffs at An Acairseid, fine grained homogenous quartz microgabbros are found as little as a metre from the poorly exposed contact. The contact zone is, however, cut by numerous aphyric microgranitic veins. These have a granular hornfelsed texture and resemble the veins developed in a similar setting on the western side of Hill 90 ( section 3.2.2 ). The quartz microgabbros pass east into homogenous aphyric olivine - bearing microgabbros, but exposure is not good enough to determine whether the transition is abrupt or gradational.

A zone of pale, intensely altered dioritic to granitic rock, 5 to 10m wide, separates the quartz gabbros from the country rocks in the cliffs northwest of An Acairseid ( Grid Ref. 43326370 ). Granitic veins are particularly abundant in the doleritic hornfelsed cone sheets which form most of the country rocks in the cliff, implying that this particular swarm of cone sheets predates coarse, unchilled ( and hence probably M1 ) granitoid rocks. However, because of the abundance of Inner Series rocks in this area, it is also possible that these granitic veins may be related to the Inner Series, as may much of the metamorphism in the area ( see below and sections 3.3 and 3.4 ).

The small area of pelitic hornfelses at the northern end of Druim na Cloise ( Grid Ref. 419654 ) was originally mapped as an isolated body within the MBG ( Richey et al. 1930 ). However, the basic rocks between it and the main area of country rocks to the south were found to be homogenous fine - grained doleritic hornfelses and are therefore considered to be a cone sheet swarm ( see Map 1 ). On the other hand the pelitic rocks around 419654 are unusual in being intensely net - veined by unchilled granitic rocks ( which also cut feldspar - phyric basic sheets in the pelites ) and in showing mesoscopic rheomorphism ( Plate 3.33 ), which is an extremely unusual phenomenon in the pelitic hornfelses around the pluton. In addition, the surrounding quartz microgabbros contain coarse granitoid lobate inclusions and granular quartzite xenoliths. These features suggest that this body of pelitic rocks may in fact be a pseudoscreen largely or wholly enclosed in the MBG; the exposures to the south are not continuous enough to exclude this possibility. The high degree of melting shown by the pelites, which correspond to the highly refractory Middle Pabba Beds, could

**Plate 3.33.** Brecciated dolerite sheet veined by rheomorphosed pelite, north end of Druim na Cloise ( grid reference 419654 approx. ). Note the contortion and disappearance of banding in the pelite close to where it cuts the refractory dolerite.

**Plate 3.34.** Pocket of recrystallised granophyre in residual quartzite block in Lower Pabba Beds, Druim na Cloise. Sample 276/2. Crossed polars, field of view 0.75mm.



be explained by such a model ( see Chapter 4 ).

#### 3.2.4.3. Country rocks and minor intrusions on Druim na Cloise.

Similar banded pelitic hornfels with green ferruginous nodules ( now composed of hornblende + epidote + sphene + feldspar + calcite or quartz, which is probably a retrogressive assemblage ) make up the bulk of the country rocks on Druim na Cloise. The pelites have a fine - grained granular texture and contain relicts of an early orthopyroxene + opaques + quartz + quartz - feldspar intergrowth ( quenched melt ? ) assemblage which is largely overprinted by a later biotite - rich assemblage ( see Chapter 5 ). On the basis of cross - cutting relationships with minor intrusions, these correspond respectively to M1 and M2 assemblages elsewhere in the contact aureole of the pluton.

The bedding in the pelites dips uniformly south - west at about 25° and as a result well - bedded quartz - rich and semipelitic hornfels equivalent to the interbedded sandstones and ferruginous siltstones of the Lower Pabba Beds are exposed close to the contact on the eastern side of Druim na Cloise. These show evidence of higher degrees of partial melting at outcrop than do the overlying pelites, in the form of disruption of the quartzite beds into rounded residual blocks like those in the rheomorphic breccias at Glendrian Bay, although the rocks retain an overall banded structure implying that little if any displacement of the blocks occurred during melting. In thin section, the rocks can be seen to have been affected by pervasive M2 metamorphism producing a biotite and hornblende bearing assemblage, although pockets of interstitial granophyre are still recognisable in the quartzite blocks ( Plate 3.34 ).

A number of groups of minor intrusions cross - cut the country rocks and the MBG on and around Druim Na Cloise and place constraints on the timing of M1 and M2 in this area. Firstly, a series of plagioclase - phyric but olivine and orthopyroxene poor dolerite cone sheets, dipping shallowly to the east, cut both the country rocks and the MBG, but not the adjacent Inner Series rocks. These dolerites, which are similar petrographically to those cutting the MBG on the western coast ( see above ), show fairly intense M2 - type alteration, with the development of green amphibole, opaques and recrystallised augite in place of primary ophitic augite. M2 in this area is therefore constrained to postdate the MBG and could well be as late as the emplacement of the Inner Series

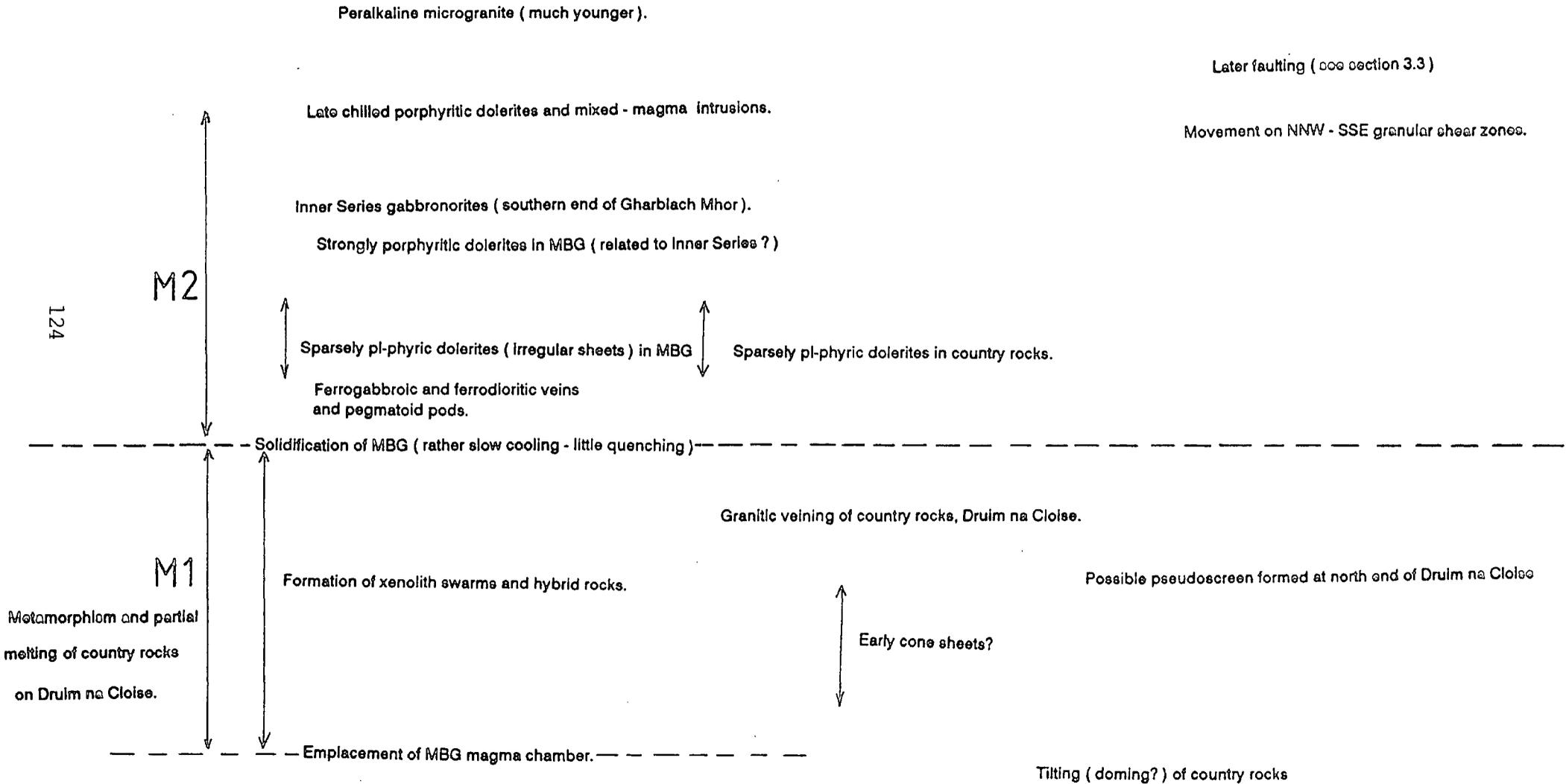
( section 3.4.6 ). A second group of minor intrusions is made up of steeply north - east dipping fine - grained porphyritic dolerites with prominent chilled margins and only slight alteration to chloritic assemblages. These therefore postdate most or all of M2 and also cut the Inner Series around An Acairseid ( Map 4 ). Thirdly, a group of hybrid pillowed dolerite - granophyre sheets, also only showing low grade chloritic alteration, also cuts both the Inner Series and the MBG. These two groups of relatively unaltered intrusions correspond to the Group 4 intrusions in the An Acairseid to Kilchoan sector of the contact ( section 3.2.5.4 ). Finally, a completely fresh sub - conformable sheet of peralkaline aegirine - hedenbergite microgranite intrudes pelites on the summit of Druim na Cloise. This is the latest intrusion in the area.

#### **3.2.4.4. Age Relationships of the MBG and associated rocks between Ardnamurchan Point and Gharblach Mhor.**

A relatively small number of samples were collected from the Ardnamurchan Point - An Acairseid sector of the MBG and aureole during the course of this study: as a result, age relationships, particularly relative to metamorphic activity, of some of the rocks are poorly constrained. However, as far as can be deduced, the same general age relationships apply in this area as on the northern margin of the intrusion, although with the following differences:

- 1). Many of the dolerite cone sheets in the Druim na Cloise area are intensely hornfelsed ( although this may simply reflect the greater intensity and higher grade of M2 in this area ), suggesting that they were affected by M1. These hornfelsed dolerites are cut by granular - textured granitic net veins whose age is uncertain ( M1 or M2? ).
- 2). A number of M2 feldspar - phyric dolerite cone sheets were emplaced into the Marginal Border Group itself, particularly along the coast at Ardnamurchan Point and Gharblach Mhor, where they were emplaced when the contact zone was still partly molten.
- 3). A number of weakly metamorphosed to fresh doleritic and felsic minor intrusions were emplaced during the latest stages of M2 or after it, some of which cut Inner Series rocks as well ( section 3.4.5 and Map 4 ).

Fig. 3.9. Age relationships in the area between Ardnamurchan Point and An Acairseid.



These age relationships are summarised in Fig. 3.9.

### **3.2.5. An Acairseid to Kilchoan.**

Although this sector of the outer contact of the MBG is both the longest and the one with the greatest topographic relief, it is also amongst the worst exposed, particularly on the north side of Druim na Gearr Leacainn ( Map 1 ). Areas of good exposure are confined to the intertidal zone and sea cliffs on the eastern side of An Acairseid, and a few small areas in the walls of the north - south trending gullies and gorges between Beinn nan Codhan and the western end of Druim na Gearr Leacainn. Nevertheless, the contact zone in this sector, and in particular its precise geometry, provides crucial evidence regarding both previous interpretations of the geometry of the intrusion ( section 1.2 ) and the alternative model presented at the end of this chapter. The model for the emplacement of the intrusion and the deformation of its wall rocks ( Chapter 7 ) and the reconstruction of magmatic processes in the boundary layer corresponding to the contact zone of the Marginal Border Group ( Chapter 4 ) also depend greatly on the field evidence presented in this section. Finally, unlike other sectors of the contact, which are close to, or coincide with, the coast, the southern margin lies up to 1.7 km from the coast. Age relationships, structures and the distribution of metamorphism in the intervening country rocks are particularly useful in the study of pre - Hypersthene Gabbro events, which is necessary for the interpretation of observations made in rocks closer to the contact where the effects of this earlier history have been heavily overprinted by the pluton.

#### **3.2.5.1 The Orientation of the Contact and the effects of Radial and Concentric faults upon it.**

M.K.Wells' inference that the outer contact of the MBG dips shallowly outwards ( Wells 1954, 1978 ) implies that its surface outcrop should be strongly influenced by the considerable topography ( up to 220m ) present in this sector of the contact. Along most of its length this does not appear to be the case. A particularly clear example occurs on the long slope leading up to Tom na Moine and Stacan Dubha from the east, where the contact runs almost perpendicular to topography for a horizontal distance of 750m and a vertical distance of 150m, between 47306425 and 46696383. Two more areas where the contact cuts across steep terrain are on the northwest side of Druim na Gearr Leacainn ( Grid Square 455634 ) and on either side of Dubh Chreag ( 453634 and 452634 ). Finally,

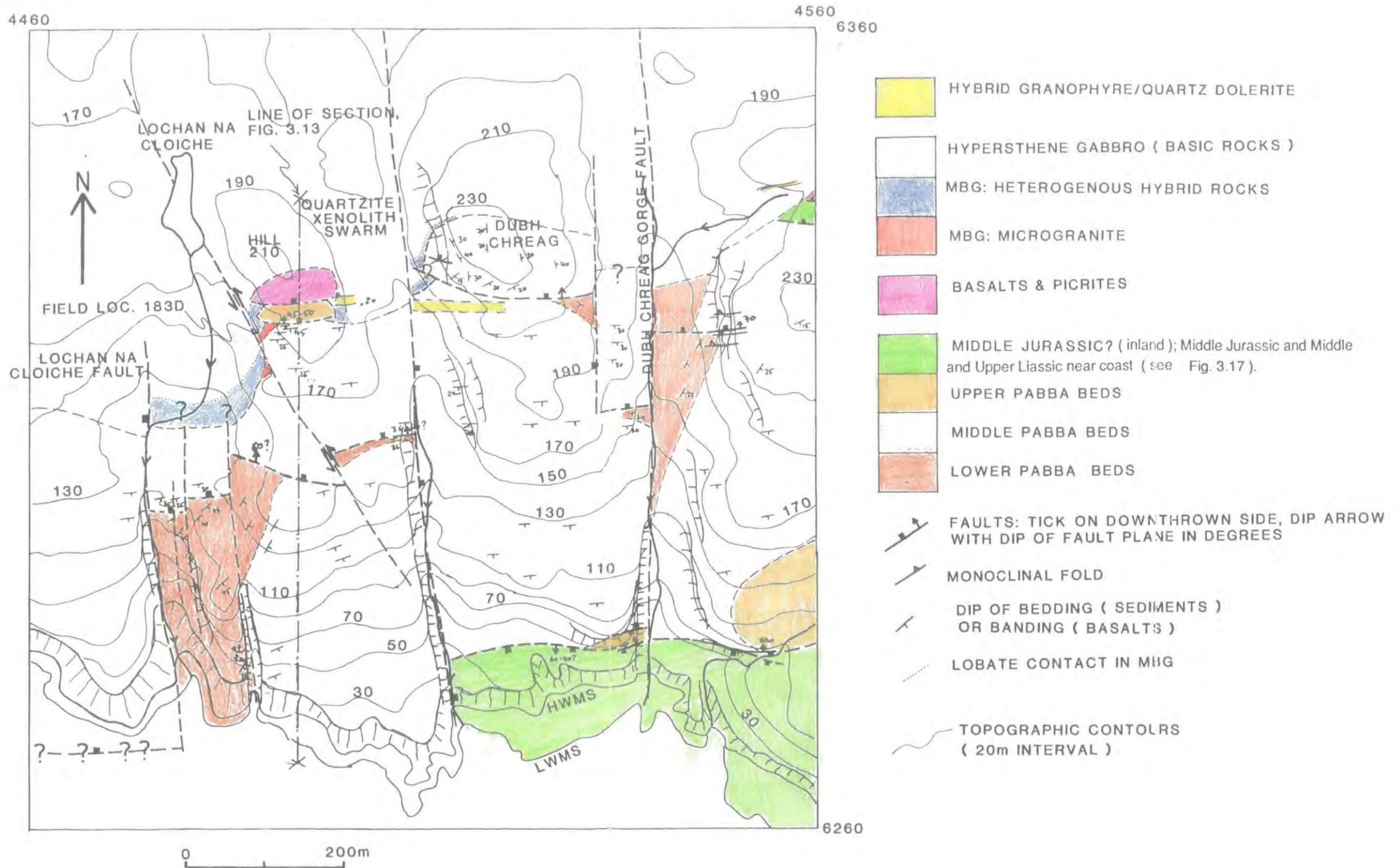
the contact exposed in the sea cliffs on the east side of An Acairseid is also very steep. However, two small areas of shallowly south - dipping contact are actually exposed, on Beinn nan Codhan ( 43206328 ) and in the eastern wall of the Dubh Chreag gorge ( about 45466328 ). The former is perhaps 10m wide, whilst the latter is about 20m wide ( measured north - south in both cases ) and appears to form a step between steeply dipping sections of the contact ( see Figs. 3.11 and 3.14 ).

The main evidence cited in support of Wells' interpretation of the contact geometry lies in the area between An Acairseid and Dubh Chreag, which forms a ridge which is dissected by a series of north - south gullies and gorges. Wells describes the contact as 'forming a V pointing southwards in each valley and curving northwards over each hill' ( Wells 1954, p.371 ). The problem with this interpretation is that it assumes the contact to be essentially planar, whereas the contact is in fact disrupted by a number of approximately radial and concentric faults.

The 'radial' faults, which in fact include both truly radial faults and a set which strikes  $140^{\circ}$  to  $150^{\circ}$  and is therefore approximately radial to the contact in this area, are all subvertical or very steeply inclined. They mainly occur in the prominent linear gullies and gorges noted above, as a result of preferential erosion of rocks weakened by the intense hydrothermal alteration haloes which occur around the faults ( see below ). The resultant poor exposure means that their age relationships and the sense of movement on them usually have to be inferred indirectly. In many cases it can be shown that these faults have long and complex histories.

Two of these faults in particular, the Lochan na Cloiche fault and the Dubh Chreag gorge fault ( see Map 1 and Fig. 3.10 for locations ), show evidence for early movement, probably predating M1 metamorphism, in the form of large - amplitude monoclinial ( drag ? ) folds which bring the bedded pelite/ferruginous siltstone sequence of the Lower Pabba Beds to the surface on the eastern side of each fault ( Fig. 3.10 ). These folds predate all the cone sheets in the area and also concentric faults which were active during M1 ( see below ). The orientation of the folds suggests that downthrow on these faults was to the east at this stage, although the outcrop pattern suggests that they have reversed throw ( i.e inverted ) during subsequent movements: there is very little offset of the top of the Lower Pabba Beds across the Dubh Chreag gorge fault whilst the Lochan na Cloiche fault actually has an overall downthrow to the west.

Fig. 3.10 Structure of the southern margin of the Hypersthene Gabbro around Dubh Chreag, showing interaction of concentric normal and radial faults, and downfaulted pseudoscreens in the Marginal Border Group. Cone sheets and dykes omitted for clarity: note that these provide important age constraints ( see text )



A number of true radial faults, particularly those on the east side of Tom na Moine ( Grid Sq. 471639 ) and between Hill 210 and Dubh Chreag ( Fig. 3.10 ) certainly seem to have been active during M1, as compartmental or transfer faults which accomodated M1 movement on the concentric faults discussed below. The clearest example of this is a relatively small fault on the east side of Dubh Chreag which terminates at a concentric fault, at 45326312, and seems to be a purely compartmental fault.

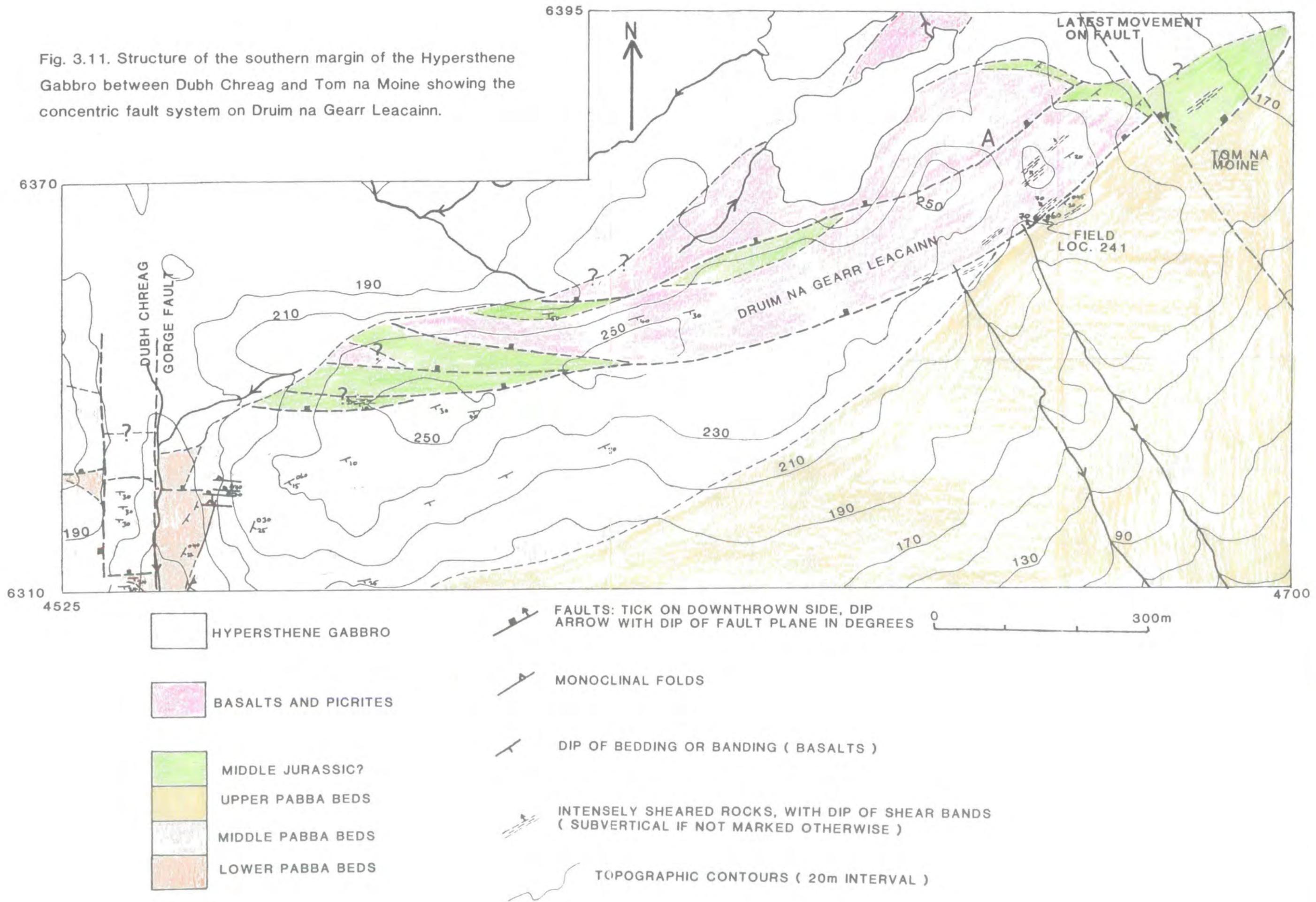
Subsequent movement on most of the radial faults is indicated by their association with zones of intense M2 hydrothermal alteration ( see 3.2.5.5 ) and by the fact that they cut the MBG, the Inner Series and, in a few cases, later intrusions as well. The faults to the west of Dubh Chreag also appear, from the distribution of linear gullies, to link up with the pyroxene - hornfels facies shear zones in the western part of the Inner Series ( sections 3.3 and 3.4.7 ). There is therefore evidence for movement on these faults during M2 and much later, postdating the Hypersthene Gabbro altogether.

The sense of movement on most of these faults is constrained to be mainly dip - slip near the contact ( see section 3.2.5.4 for discussion of movements away from the contact ), from offsets of outward - dipping sedimentary bedding and inward - dipping cone sheets and faults. However, later movement on two faults bearing about  $140^{\circ}$  ( parallel to the regional fault trend ( see Chapter 7 ) ), the Tom na Moine and Hill 210 faults, appears to be sinistral strike - slip.

The concentric faults referred to above are characterised by extensional geometries, with both inward dips and inward downthrows. They have curved traces at outcrop which are roughly concentric to the pluton but tend to have tighter curvatures than the contact of the Marginal Border Group itself. This gives the hanging - wall blocks a crescent - shaped geometry in plan ( see Fig. 3.11 ), except where they are truncated by radial transfer faults.

The largest and most complex group of concentric faults occurs on Druim na Gearr Leacainn ( Map 1 & Fig.3.11 ), where calcareous sediments ( middle Jurassic? ) and Tertiary alkali basalts and picrites are downfaulted against the Middle and Upper Pabba Beds. Although numerous minor faults and cataclastic shear zones are exposed in the hanging - wall block, the main faults parallel to these are very rarely exposed. The most important exposure is at Grid Ref. 46646365 ( field loc. 241 ), where calcareous, shell debris bearing pelites of the Upper Pabba Beds are separated

Fig. 3.11. Structure of the southern margin of the Hypersthene Gabbro between Dubh Chreag and Tom na Moine showing the concentric fault system on Drum na Gearr Leacainn.



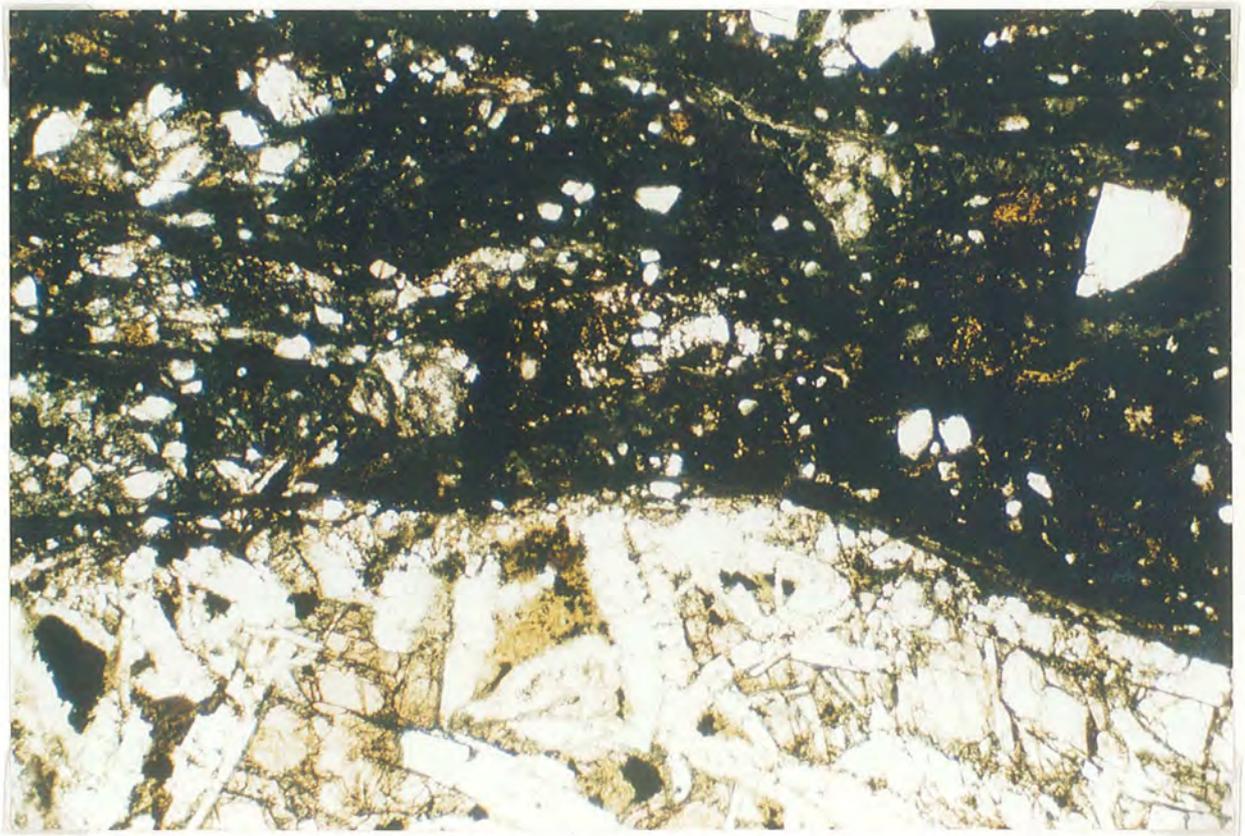
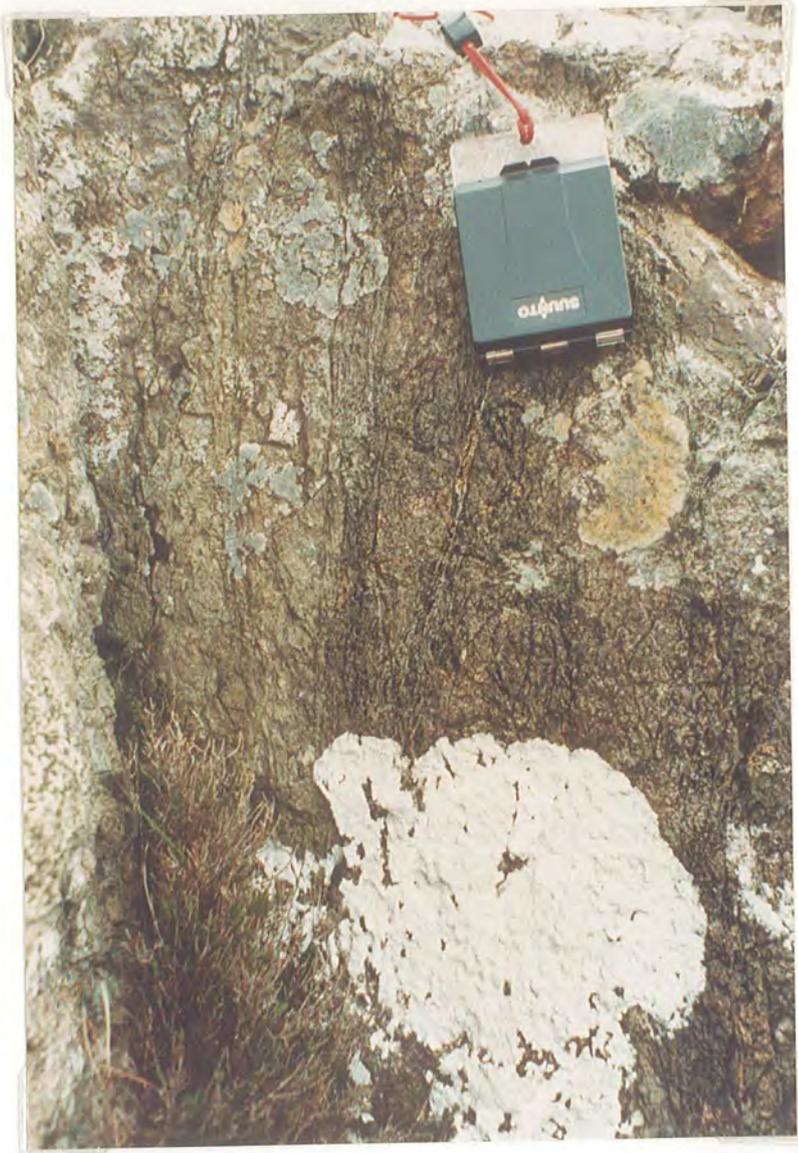
CONE SHEETS OMITTED FOR CLARITY

from hornfelsed basalts by a zone of intensely sheared rock about 5m wide ( Plate 3.35 ). Within this zone, the most intense deformation is found in narrow banded seams spaced at intervals of a few millimetres to a few centimetres: in thin section these are composed of fine - grained cataclastic material with numerous rounded clasts of olivine, titaniferous augite and plagioclase from the host basalt or picrite ( Plate 3.36 ). The brittle style of deformation in these rocks, together with the lack of annealing, suggests that deformation took place at relatively low temperatures, after the high - grade metamorphism ( M1 ) in the area. At this outcrop the fault zone is cut by a completely undeformed and unmetamorphosed basaltic sheet and by an earlier ultraporphyritic, near - anorthositic coarse dolerite cone sheet. This latter contains a few irregular brittle microfaults, with epidote deposited in cavities generated by the fault movements, and shows weakly developed low temperature alteration to epidote and sericite. In the subdivision of the cone sheets adopted in section 3.2.5.4 below, it is a Group 3 cone sheet. This indicates that movement on this fault mainly occurred prior to the latter half of M2 in this area and had entirely ceased by its end, from the age relationships of the cone sheets to the metamorphism ( section 3.2.5.5 ). The dominant set of shear bands in the fault zone strike  $070^{\circ}$  and dip northwest at  $80$  to  $90^{\circ}$  overall but anastomose. A second set strike  $090^{\circ}$  and is subvertical to very steeply south - dipping. These variously cut and are cut by the main set of shears and may represent a conjugate set, implying a component of sinistral strike - slip on the fault or east - west compression concentric to the margin of the pluton ( see Chapter 7 ). Within areas of low strain, rotated early shear bands and hornfelsed ( M1? ) microgranitic veins indicate downthrow to the north in adjacent intensely deformed zones, consistent with the map - scale evidence ( Fig. 3.11 ).

An adjacent outcrop, just north of the main fault zone at 46526368, also shows evidence for M2 movement on the faults, in the form of a weakly altered vein of feldspar - phyric basaltic rock injected up an irregular shear zone and showing intense deformation of its chilled margins, indicating syn - deformational emplacement during M2. In contrast, minor subvertical shear zones up to a centimetre wide, in quartzite hornfelses towards the western end of the summit ridge of Druim na Gearr Leacainn ( 45666336 ) have a coarse annealed, originally cataclastic texture which may therefore have formed during M1. The cataclastic texture would in this case indicate an extremely high strain rate. A final piece of evidence for the age of movement on these faults is provided by outcrops of MBG rocks adjacent to the intersections of some of the faults with the contact, in grid squares 466638 and 456634 in particular. **These are not cut by the faults.** This implies that most if not all of the

**Plate 3.35.** Zone of intensely deformed fault rock separating hornfelsed basalts to north ( left of plate ) from calc - pelites of the Upper Pabba Beds. Field location 241, on main fault at the southern side of the downfaulted block which forms the summit ridge of Druim na Gearr Leacainn ( Grid reference 46646365 ). The subvertical fabric is formed by parallel zones of particularly intense deformation ( see Plate 3.36 ).

**Plate 3.36.** Cataclastic shear zone cutting alkali dolerite from northern half of fault in Plate 3.35 ( sample 241/2 ). Clasts within cataclasite are fresh calci plagioclase and titanite. The plate is overexposed to reveal anastomosing shear zones within the main shear zone. Some of these are irregular and vein - like; they may be pseudotachylitic. Plane polarised light, field of view 3.5mm.



movement on these particular faults took place during M1 or earlier. It should be noted that the appropriate outcrops are absent adjacent to the intersection of the main fault between the basalts and the Pabba Beds with the contact, and it is this fault for which there is outcrop evidence for M2 movement, as described above.

Whatever the age of the faults on Druim na Gearr Leacainn, extrapolation downwards and southwards of the unconformity beneath the basalts, which outcrops locally on the north side of Druim na Gearr Leacainn, and addition of the 70 - 100m of sediments eroded from above the Middle Pabba Beds to the south of the fault, leads to an estimate of the total throw on the concentric faults of 130 to 250m. This is exclusive of downthrow on the fault running from 466638 to 459635, which increases markedly from east to west.

A number of similar concentric faults occur to the west of the Dubh Chreag Gorge fault. These fall into two groups divided by the radial Lochan na Cloiche fault ( see Fig. 3.10 ). To the west of this fault, the top of the bedded ferruginous siltstone - pelite sequence of the Lower Pabba Beds is exposed close to sea level at An Acairseid but not at all close to the contact on Beinn nan Codhan, which is formed by Middle Pabba aluminous pelites. To the east of the fault, Lower Pabba lithologies outcrop at structurally higher levels, in the footwall blocks of a number of northward - downthrowing concentric faults on the southern sides of Hill 210 and Dubh Chreag itself. This suggests that the Lochan na Cloiche fault has acted as a transfer fault, shifting 50 to 80m of the displacement on these faults onto a hypothetical concentric normal fault or faults well to the south. A group of small concentric inward - dipping normal faults, with dips of 40 - 70°, occur in a cliff east of An Acairseid ( Grid Ref. 43616310 ) and may be associated with the main fault to the south. These are cut by hornfelsed aphyric dolerite sheets, implying an M1 or earliest M2 age of faulting.

Three large concentric normal faults are present to the east of the Lochan na Cloiche fault. The first two are close to the summit of Hill 210 and actually displace the contact at the present level of exposure ( see Fig. 3.10 ). The northernmost fault is not exposed, and is inferred from the juxtaposition of basaltic and metasedimentary hornfelses, but that to the south is exposed at its junction with the MBG contact in a key outcrop at 44926325 ( field loc. 183D ). **At this point the fault displaces the contact but does not deform the MBG rocks, and is itself cut and veined by them. This indicates that it displaced the wall of the magma chamber**

corresponding to the Marginal Border Group whilst this was still molten and fluid. In a sense, the fault can be said to root out into the magma chamber. The outcrop, shown in Plate 3.37 and Fig. 3.12, contains an irregular exposure of hypersthene- and plagioclase - phyrlic, possibly cordierite - bearing microgranodiorite between partially melted and disrupted hornfelses to the north, which show abundant evidence for macroscopic mobility of high - degree partial melts, and Middle Pabba Beds pelitic hornfelses to the south. The latter dip southwards overall but contain a large number of south - verging minor folds, with axial planes dipping  $40 - 50^\circ$  north, and a few centimetre - scale south - directed M1 thrusts. These are cut by a deformed and intensely hornfelsed dolerite inclined sheet which is itself cut by buchitic veins and by the contact of the pluton itself, and must therefore be M1 or earlier in age. The plastic nature of the earlier deformation in the pelitic hornfels suggests an M1 age for the dolerite.

The contact between the pelitic hornfelses and the microgranite is irregular and lobate in places, as a result of rheomorphism in the pelite at the contact, **but is always sharp** ( Plate 3.38 ). This indicates that the microgranite did not form in situ but was emplaced during M1. In contrast the contact on the northern side of the microgranodiorite outcrop is gradational, with a transition zone of rheomorphic breccia and homogenous grey quartzofeldspathic felsite between the hypersthene - phyrlic microgranodiorite and the laminated hornfels to the north. This pattern is very similar to that seen on the northern margin of the Hypersthene Gabbro ( section 3.2.3 ), where the outer contact is sharp but the pseudoscreen at Duin Bhain has an intricately veined and gradational contact with the surrounding contact zone rocks. The fault blocks to the north of Loc. 183D, if they were eroded to a slightly deeper level than that seen at present, would have a similar geometry to the pseudoscreen at Duin Bhain. Conversely, downward displacement along an inward dipping fault would produce the outcrop pattern of Jurassic lithologies observed at Duin Bhain ( section 3.2.3 ). It therefore appears that the downfaulted blocks on Hill 210 and the pseudoscreen at Duin Bhain are examples of the same type of structure, exposed at different structural levels.

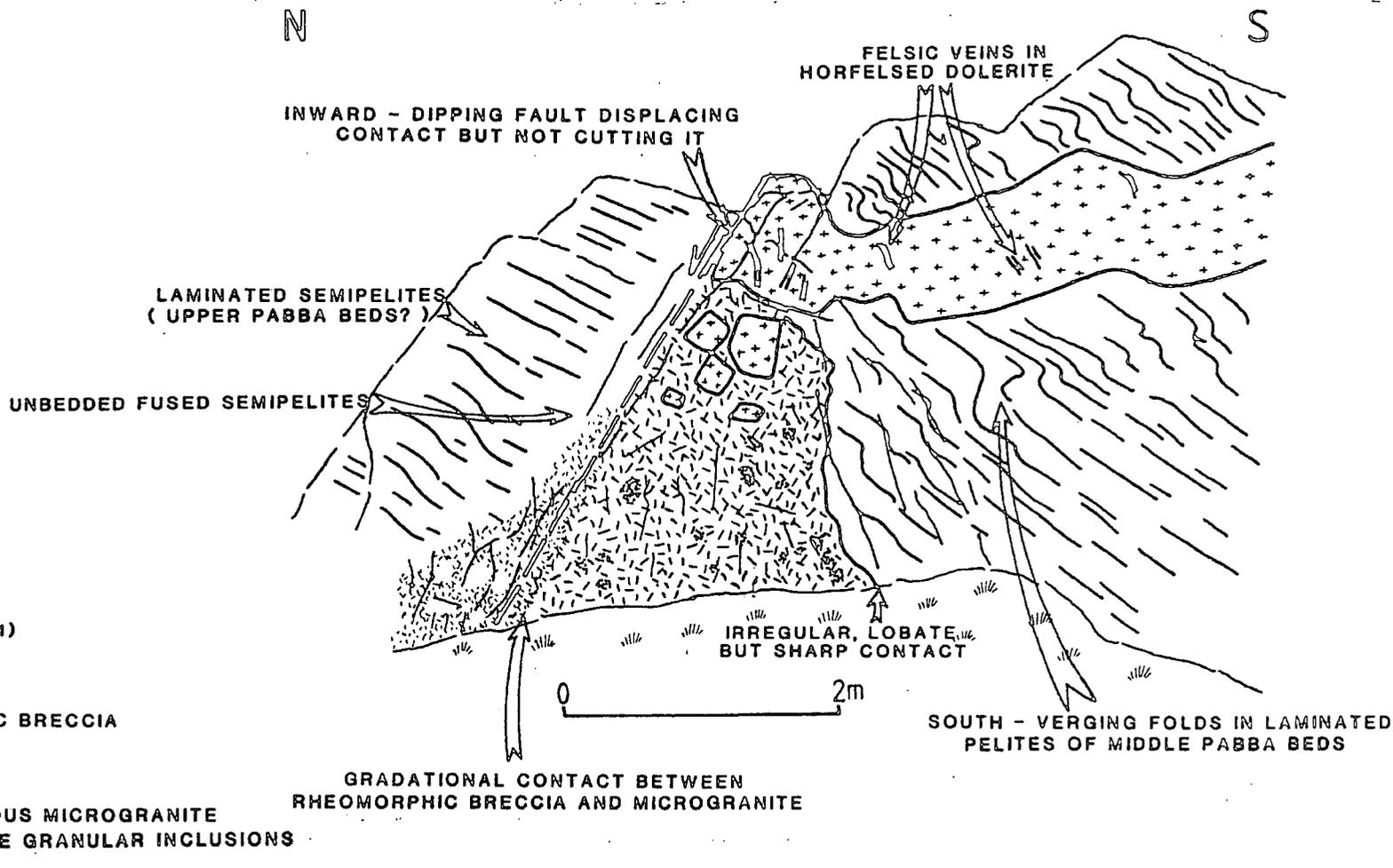
The third concentric fault on the southern side of Hill 210 is only exposed in badly weathered outcrops around 45106314, where it dips  $50 - 60^\circ$  north and is cut by weakly altered coarse feldspar - phyrlic dolerite cone sheets ( of cone sheet group 3 - see section 3.2.5.4 ): it appears to be of early M2 age at the latest. Another Group 3 cone sheet cuts the expected line of the concealed fault around 44886306. The fault downthrows Middle Pabba Beds against well - bedded ferruginous

**Plate 3.37.** Field location 183D, grid reference 44926325. The orthopyroxene - phytic microgranodiorite forms the weathered - out area of pale rock between the downfaulted block to the north ( left side of the plate ) and the main part of the wall of the MBG.

**Plate 3.38.** Field location 183D. View of the irregular but sharp contact between the microgranitoid and the Middle Pabba Beds pelites. This runs just above the pencil. Note small - scale folding of the pelitic hornfelses.



FIG. 3.12 LINE DRAWING BASED ON FIELD SKETCHES AND PHOTOGRAPHS OF LOC. 183D, SHOWING ORTHOPYROXENE MICROGRANITE BETWEEN THE MAIN WALL OF THE HYPERSTHENE GABBRO AND DOWNFAULTED BLOCK TO THE NORTH.

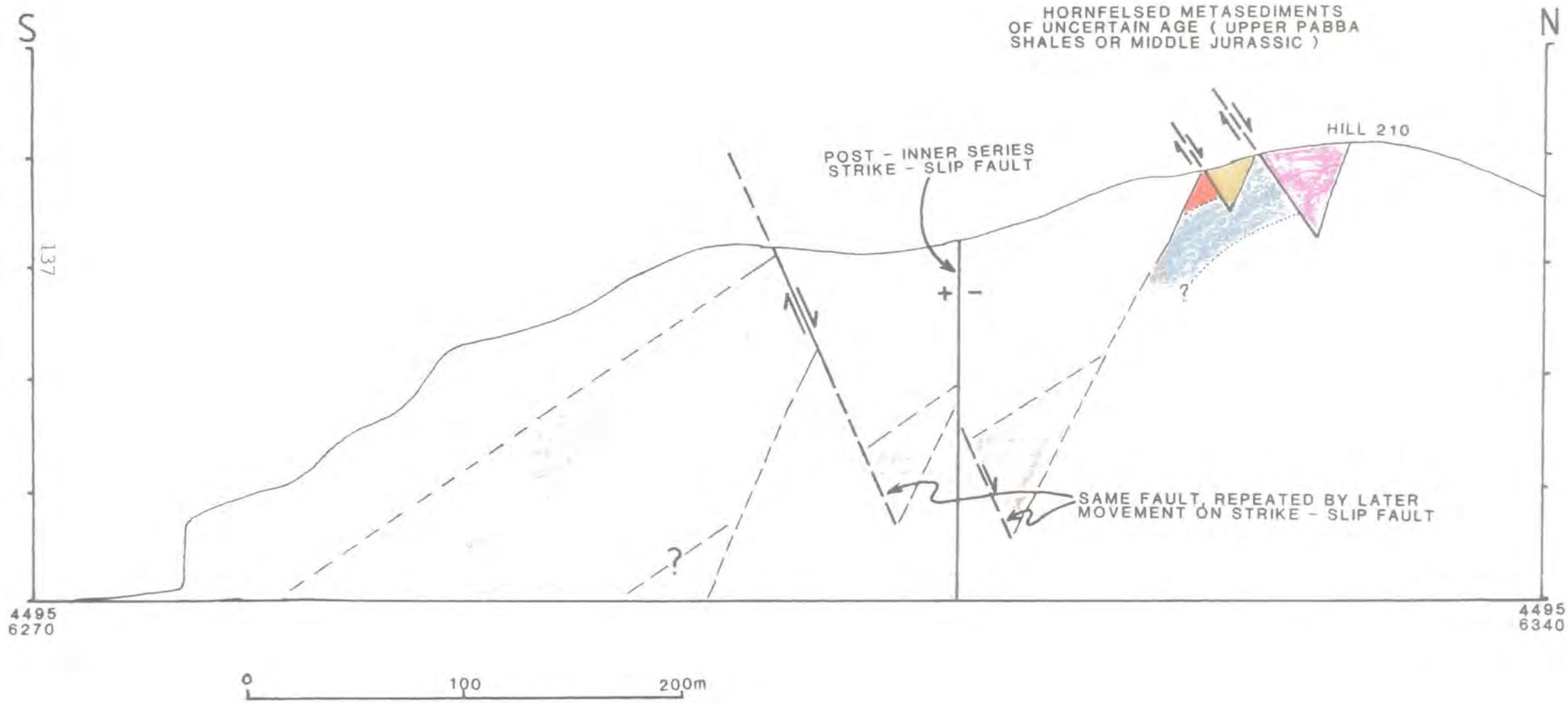


135

quartzite and pelitic hornfels of the Lower Pabba Beds and has a throw of between 70 and about 100m, from the geometrical constraints shown in Fig. 3.13: the total thickness of cone sheets along the line of section, which are not shown on Fig. 3.13, is small ( see Map 1 ). It should also be noted that the subsurface part of Fig. 3.13 is constructed on the assumption that any tectonic or thermal erosion of the hanging - wall blocks that may have taken place has not modified the orientation of the displaced contacts: as will be seen in Chapter 4, this is probably not valid. One feature of the outcrop pattern in Fig. 3.10 that suggests that such erosion may have taken place is the small size of the downfaulted block of metabasalts on Hill 210 around Grid Sq. 449632. This is much smaller than would be expected from the stratigraphic offset on the fault to the south, implying removal of material from its northern side. A similar, unexpectedly small northerly offset of the outward - dipping contact of the Hypersthene Gabbro is apparent in the hanging wall block of the poorly exposed concentric fault at the eastern end of the An Acairseid - Kilchoan sector ( Grid Sq. 4764 ), suggesting that this too may have formed when the MBG was still a magma chamber, ( i.e. during M1 ).

Additional concentric normal faults, further from the contact, may be inferred from the outcrop patterns of the units within the Pabba Beds, as defined in section 2.3. In addition to the faults offshore to the south of Beinn nan Codhan, inferred above, similar faults may be present in the poorly exposed ground south - east of Tom na Moine. The main evidence for this is Richey et al.'s conclusion that the outcrops at the eastern end of the shore section along the north side of Kilchoan Bay belong to a unit in their succession ( Richey et al. 1930; see Fig. 2.3 ) equivalent to the Middle Pabba Beds of Chapter 2. The outcrops concerned are certainly non - fossiliferous, but are rather more heterogenous, with sandy and calcareous siltstones as well as pelites, than the Middle Pabba Beds as exposed inland. If Richey et al.'s identification of these beds is correct, the occurrence of Upper Pabba Beds to the north and to the west of a NNW - SSE trending fault on the shore around 476634, suggests that a concentric inward - downthrowing fault **may** be present in the poorly exposed ground between 476635 and 480640. To the west, movement on the fault must be transferred offshore along the north - south fault at 476634. The orientation of the fault between 476635 and 480640, and the age of movement on it, are unknown, except that it does not obviously cut cone sheets in the area. This suggests that, if present, it must predate the cone sheets and therefore be M1 or earliest M2 in age ( see below for discussion of the age of these cone sheets ).

Fig. 3.13. South - North section through Hill 210 showing the effect of concentric normal faulting on the outer contact of the Marginal Border Group. No vertical exaggeration.  
 Key as for Fig 3.10. Cone Sheets omitted for clarity.



In conclusion, the disruption of the wall of the Marginal Border Group of the Hypersthene Gabbro described in this section can be largely accounted for in terms of movement on a linked system of concentric normal faults, which had the effect of extending and flattening the contact overall, as well as giving it a saw - tooth profile in detail, and radial faults which acted in part as transfer faults. This system of faults was active during M1 and the earlier part of M2. In addition, the radial faults, and faults with the regional 140 - 150° strike, were active at both earlier and later times. The consequences for Wells' interpretation of the geometry of the contact are apparent from Fig. 3.10 in particular. Around Dubh Chreag and Hill 210, the intersections of the concentric faults with the contact of the pluton occur at about the present level of exposure. The outward dipping contact is displaced northwards in the hanging - wall blocks, which are exposed on the ridges, relative to the contact in the footwalls, which are exposed in the valleys. If this effect is not allowed for, the contact appears to be conformable to the domed bedding in the country rocks when in fact it is much more steeply inclined and markedly discordant to the sedimentary bedding.

#### **3.2.5.2. Contact Zone Rocks along the Southern Margin of the Hypersthene Gabbro.**

The hypersthene - cordierite microgranodiorite at Loc. 183D is at one of a number of mostly small localities along the southern margin of the pluton at which contact zone suite felsic, iron rich intermediate and hybrid rocks are exposed. Such rocks appear to be less common along this southern margin than in the north. The commonest rock type along the contact is homogenous microgabbro, usually quartz - bearing and with rare felsic veins and irregular lobate inclusions.

At the extreme western end of the sector, the coastal section on the east side of An Acairseid ( Grid Squares 435631 to 436629 ) provides an unusually complete section through the contact zone of the MBG. The rocks present along this coastal section have however been intensely recrystallised and hydrothermally altered, probably due to the proximity of the radial and concentric fault systems in the An Acairseid area, which appear to have acted as fluid flow conduits ( section 3.4.6 and Chapter 5 ). At the northern end of the section, the rocks closest to the boundary between the Inner Series and the MBG are homogenous olivine - orthopyroxene microgabbros. Around 43556318 these contain a prominent and very dense xenolith swarm ( Plate 3.39 ). Like other swarms, this is dominated by a single lithology, in this case cordierite + orthopyroxene + plagioclase + magnetite granular hornfels with well - developed contorted banding. These are cut by coarse opx - rich cordierite norite veins, (

**Plate 3.39.** Contorted banded pelitic xenoliths in xenolith swarm within the MBG rocks on the south east side of An Acairseid. Field location 274, grid reference 43556319.

**Plate 3.40.** Igneous - textured cordierite - norite vein within pelitic xenolith from location 274 ( sample 274/3 ). Note the diffuse contact between the vein and the granular cordierite - plagioclase - orthopyroxene host rock. Plane polarised light, field of view 14mm.

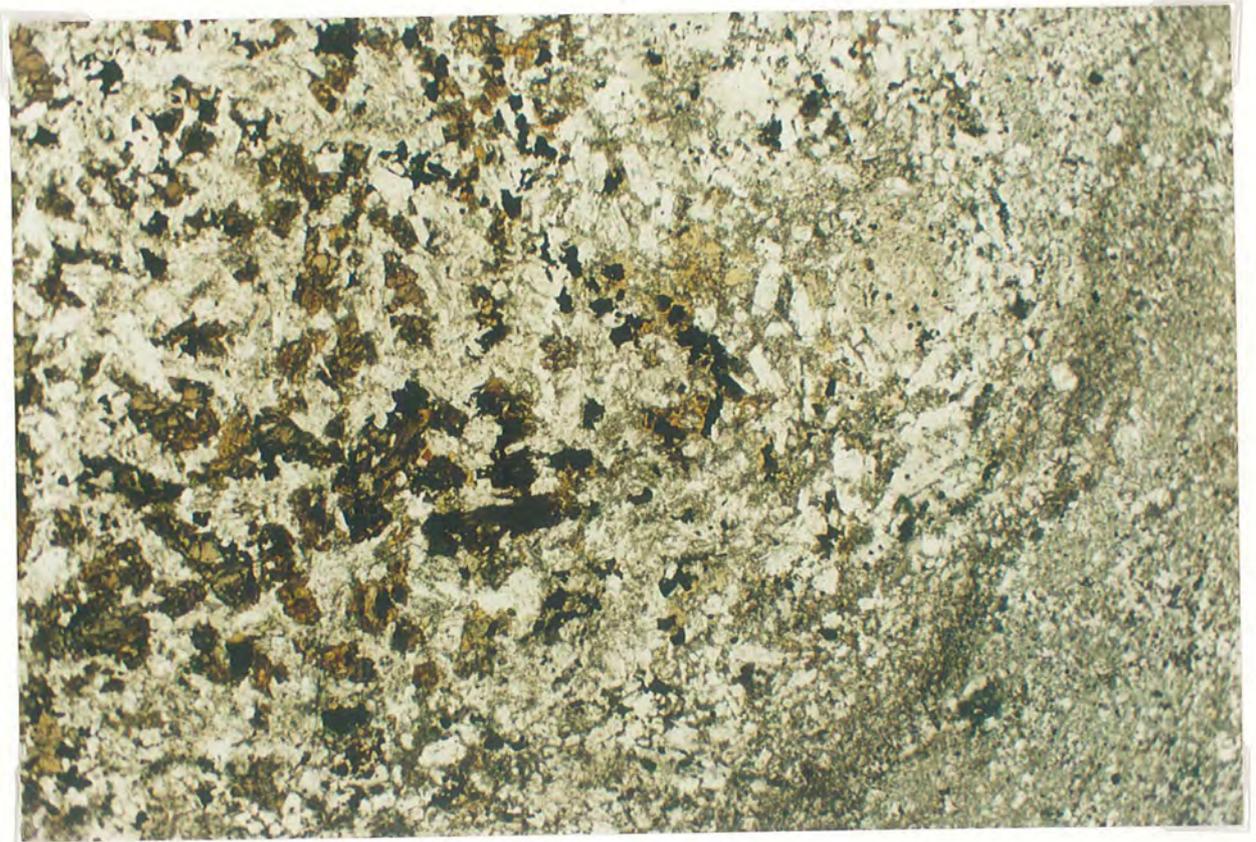


Plate 3.40 ) which on phase equilibrium grounds ( Gribble & O'Hara 1967 ) must have been produced by melting of cordierite - bearing rock rather than by infiltration of magma from the gabbroic host. Other, volumetrically minor components of this xenolith swarm include granulitic textured olivine microleucogabbro and coarser granulitic leucogabbros.

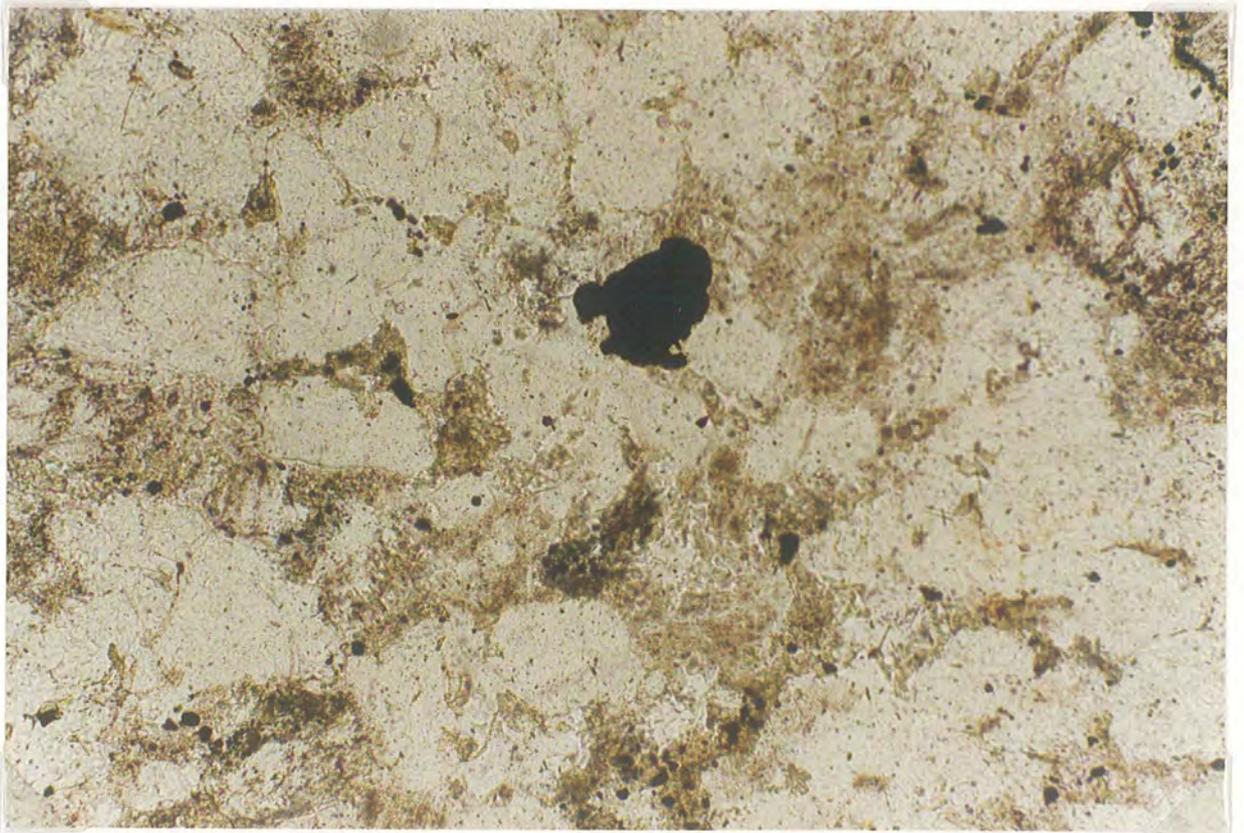
To the southeast, a 30m wide zone of mottled hybrids, ranging from gabbroic through dioritic to granitoid rocks, with extensive veining of basic rocks by more felsic rocks, passes into a 10m wide zone of relatively homogenous dolerite. This contains granitic veins and a number of granular quartzite xenoliths, which occur in clusters up to a few metres across. The homogenous, sparsely feldspar - phytic dolerite passes southeast, towards the contact, into a pillowed and backveined dolerite - microgranite complex. This contains well - developed chilled porphyritic dolerite pillows in a microgranite host with abundant rounded quartzose xenoliths ( Plate 3.41 ). The xenoliths vary from possible diopside - quartzites through grey quartzites to banded cordierite - orthopyroxene bearing xenoliths similar to those to the northwest. The southeastern edge of the pillowed rocks also forms the outer contact of the MBG at this point. Overall, the sequence of contact zone rocks at An Acairseid is very similar to those which occur at those parts of the northern margin where the contact zone suite is well developed, except that the finer - grained rocks in particular are much more heavily altered. Rather granular textures are especially common in the felsic rocks and if quench textures were ever present in these rocks they would probably not have been preserved.

The same intense overprinting of earlier igneous and/or high grade metamorphic textures is present in the country rocks in the coastal section at An Acairseid, most notably in the case of a body of rheomorphic breccia around 43616302. In the field this has the lumpy appearance characteristic of rheomorphic breccias, with randomly distributed rounded inclusions of quartzite in a grey fine - grained homogenous matrix, but in thin section the latter is a granular quartzofeldspathic rock with accessory epidote and chlorite ( Plate 3.42 ). Again, any quench textures that might have been present would probably have been obliterated by the intense hydrothermal metamorphism which subsequently affected this rock.

Between An Acairseid and the valley west of Hill 210 ( Grid Sq. 447631 ) the Marginal Border Group is dominated by quartz microgabbros with only a few lobate inclusions of coarse leucocratic rock and mottled patches. A few more mafic pillow - like bodies are present around 43846332;

**Plate 3.41.** Quartzose xenolith in pillowed hybrid dolerite - microgranitoid rock, close to the outer contact of the Marginal Border Group at An Acairseid. Field location 272, Grid reference 43586313.

**Plate 3.42.** Granular epidote and chlorite - bearing quartzofeldspathic matrix of intensely altered rheomorphic breccia at An Acairseid. Sample 267/3. Plane polarised light, field of view 1.5mm.

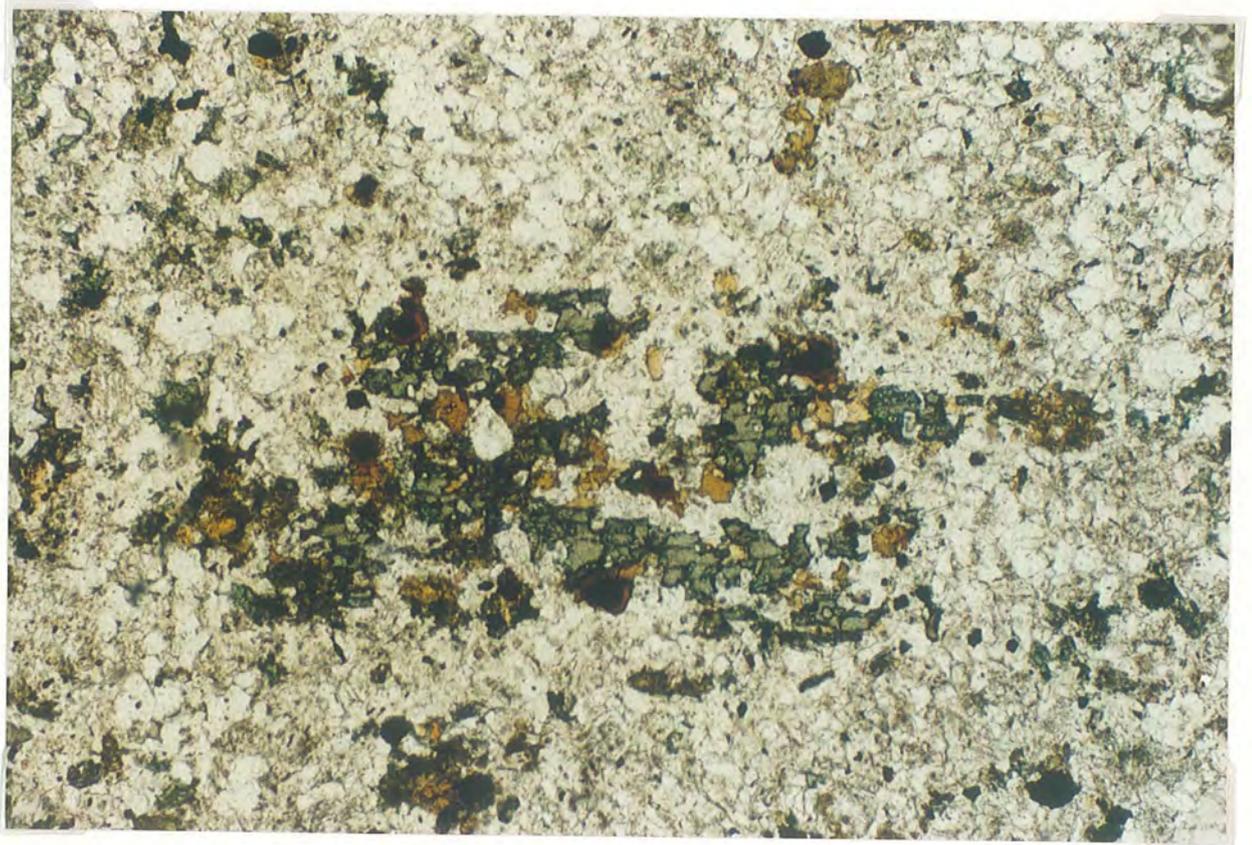
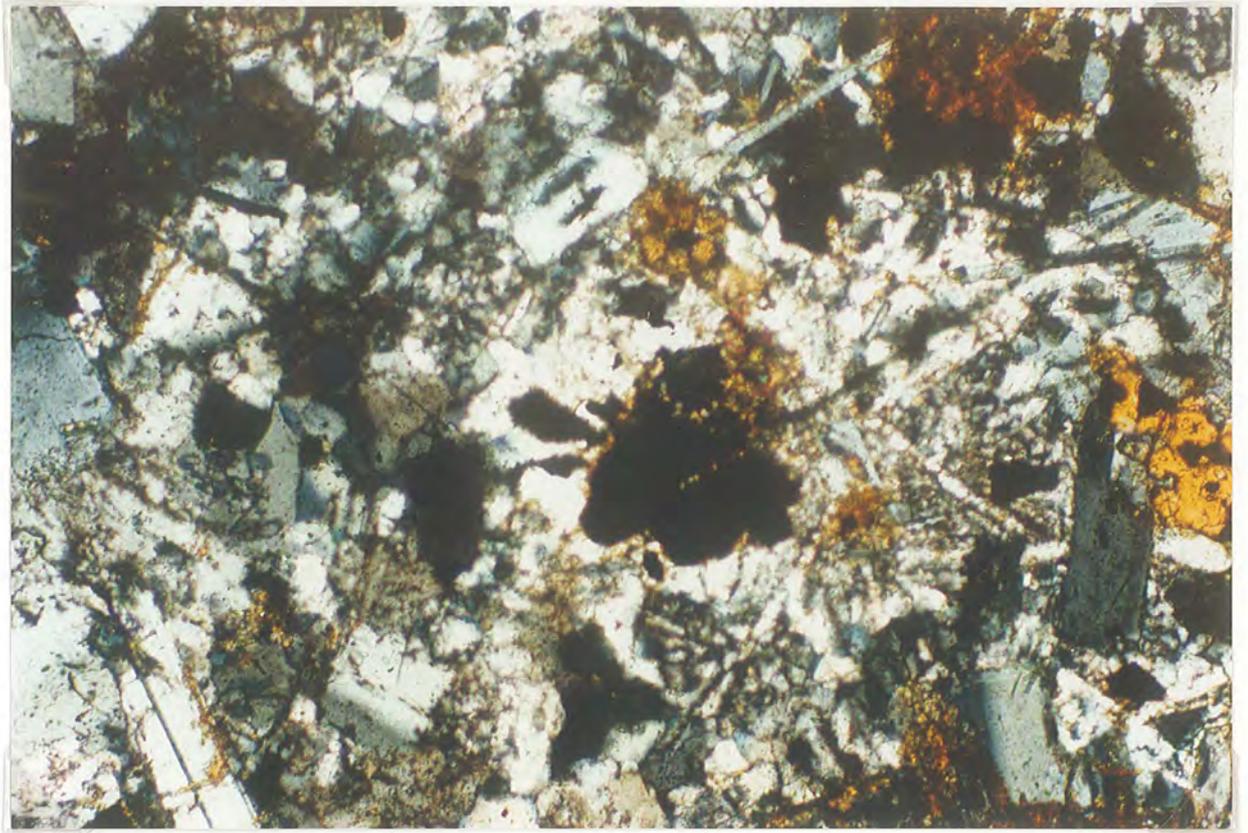


diffuse granitic veins are present in the same area and around 43976340. At the only outcrop of the contact itself ( 43846325 ), the dominant MBG rock type is a pale quartz diorite or granodiorite, containing rounded microgabbro inclusions, which is cut by a network of granitic veins which do not, however, cut the country rocks. Two small xenolith swarms occur, respectively of quartzose hornfels ( 43876328 ) and of banded ferruginous hornfels similar to the cordierite norite xenoliths at An Acairseid, at 44276340.

In contrast, the area south of Hill 210 is notable for the presence of large amounts of felsic, hybrid and strongly fractionated rocks within the MBG, and especially for two rock types not found elsewhere around the Hypersthene Gabbro. One of these is the hypersthene - cordierite microgranite exposed at Loc. 183D. In detail, this rock is extremely heterogenous, with up to 10% of the rock being composed of centimetre sized rounded inclusions. These are mainly fine - grained granular hornfelses, with the assemblages cordierite + orthopyroxene + opaques + plagioclase + K-feldspar + quartz, cordierite + opaques + plagioclase + quartz and cordierite + opaques + hercynite + plag. + K-feldspar. K-feldspar and quartz typically occur as poikiloblastic intergrowths in these inclusions and may represent a crystallised interstitial melt. Rarer inclusions also present include fine - grained mafic xenoliths and a coarser, partly igneous - textured inclusion. The latter contains zoned euhedral plagioclase phenocrysts, sub - poikilitic orthopyroxene, euhedral opaques, and possibly poikiloblastic cordierite ( augite is definitely absent ) set in a matrix of quenched hollow plagioclase laths and quartz - feldspar microgranophyre ( Plate 3.43 ). This groundmass texture is particularly significant because it indicates rapid quenching of the partially molten inclusion and of its host magma. Textural evidence of this quenching is not preserved in the microgranite itself because of subsolidus recrystallisation ( see below ). The microgranite contains prominent poikilitic ( or poikiloblastic? ) to skeletal orthopyroxene in some sections ( Plate 3.44 ), together with unzoned but rounded andesine plagioclase porphyrocrysts and glomerocrysts, and a few probable cordierite porphyrocrysts ( positive identification of these is difficult because of possible confusion with plagioclase; no cordierite was found in the probe section of this rock analysed by electron microprobe). The groundmass is heterogenous, with quartz - rich and feldspar - rich areas a few millimetres across, and is intensely altered. Secondary biotite and blue - green ferrohastingsitic hornblende are present but most of the mafic phases have been replaced by chlorite and the feldspars are often very turbid. Most of the groundmass has a granular texture, although in places irregular quartz - alkali feldspar intergrowths are present ( Plate 3.45 ). These may be relict after an original microgranophyric tex-

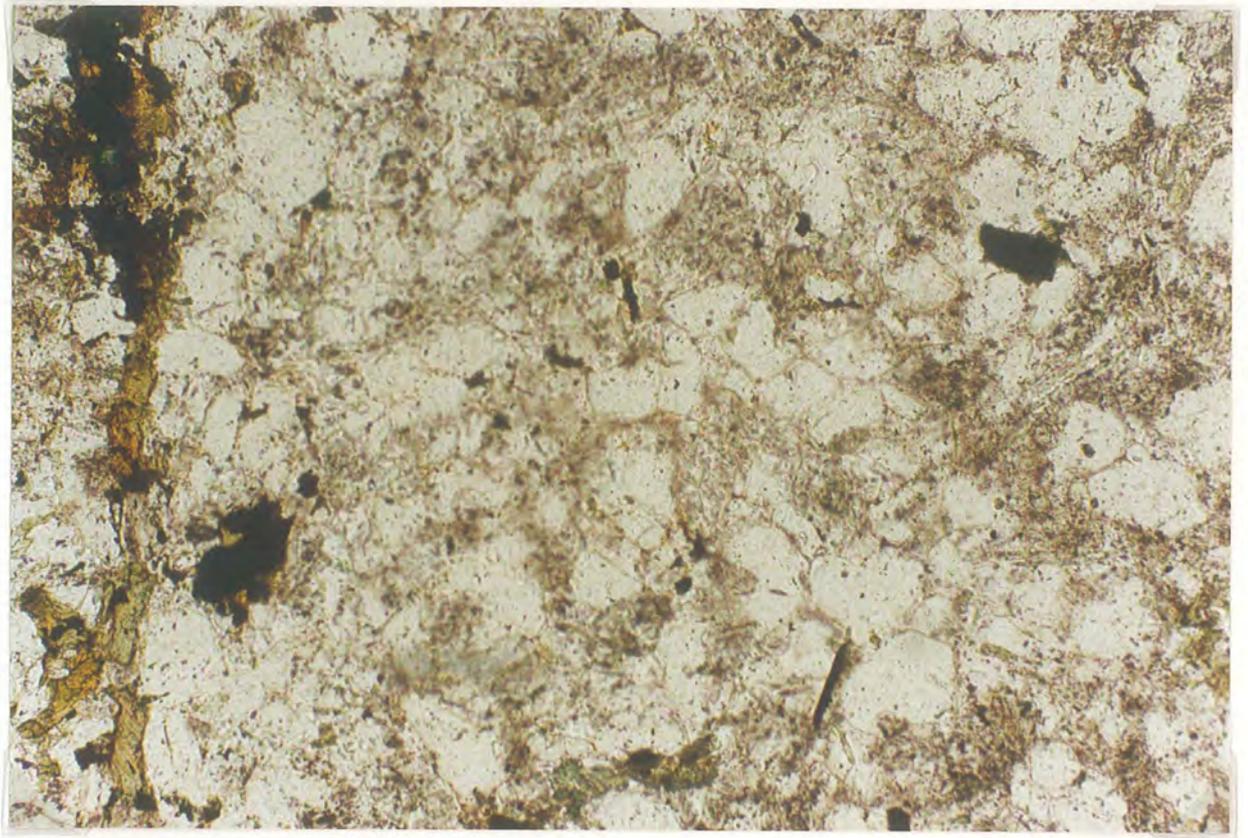
**Plate 3.43.** Quench - textured orthopyroxene - bearing inclusion in recrystallised orthopyroxene microgranitoid, sample 183D1. Plane polarised light, field of view 1.5mm.

**Plate 3.44.** Skeletal orthopyroxene poikilocryst in recrystallised orthopyroxene microgranitoid, sample 183D5. Plane polarised light, field of view 3.5mm.



**Plate 3.45.** Recrystallised groundmass of orthopyroxene microgranitoid, showing possible relict interstitial granophyre. Note amphibole - filled tensile fracture ( hydrothermal vein ) at extreme left of field of view. Sample 183D5. Plane polarised light, field of view 1.5mm.

**Plate 3.46.** Leucocratic veins and lobate inclusions in quartz ferrodiorite on south west side of Hill 210, field location 183B1. Grid reference 44906325.



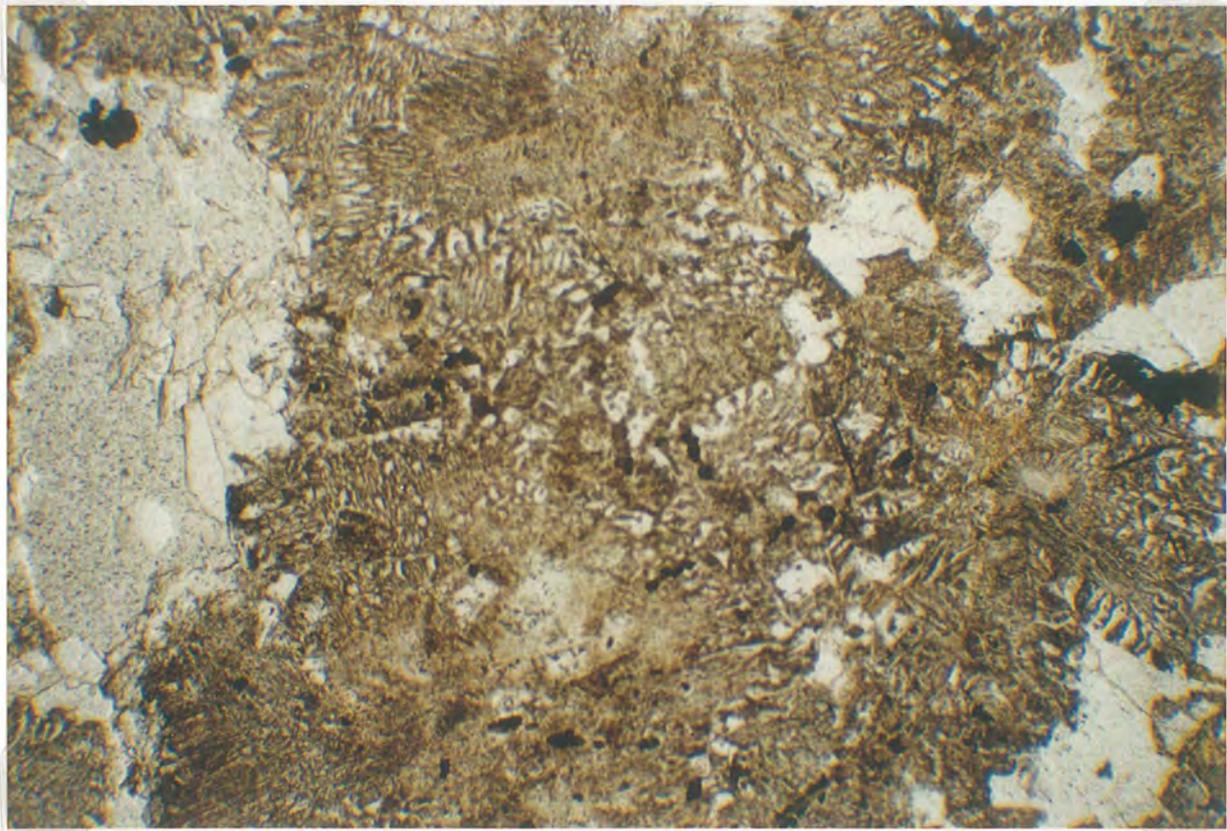
ture. Mirolitic cavities lined with drusiform quartz and filled with fibrous chlorite are also present. In general, however, the groundmass does not have an igneous texture although the appearance of the rock at outcrop and its intrusive relationships ( Plates 3.37 and 3.38, above ) clearly show that it is igneous in origin. The intense alteration of the rock as a whole suggests that the groundmass may represent the final granular - textured product of the progressive devitrification and alteration of a quenched glass which was postulated by Lofgren ( 1971 ). The glass in the groundmass of this rock implies that, as on the northern margin of the pluton, hydrothermal alteration during M2 metamorphism was preceded by an abrupt quenching event which affected the contact zone of the MBG. As noted above, the microgranite itself has an unchilled contact against the country rocks to the south and grades into a rheomorphic breccia in the downfaulted pseudoscreen to the north, implying that it was emplaced during M1.

The second unusual rock type on the side of Hill 210 is a quartz ferrodiorite. In contrast to the ferrodiorites and ferrogabbros found elsewhere in the contact zone of the pluton this rock forms a continuous, almost homogenous body on the western side of the hill, and appears to be M1 rather than M2 in age. Several outcrops contain coarse leucocratic lobate inclusions and irregular coarse granitic veins ( Plate 3.46 ). Similar but very intensely altered rocks form the coarse mesocratic component of a conmingled diorite - porphyritic dolerite rock found in outcrops on the eastern side of Hill 210.

Although the downfaulted blocks on Dubh Chreag and Druim na Gearr Leacainn are larger than those on Hill 210, the volumes of acid and hybrid contact zone rocks associated with them are much smaller. The same is also true of the very poorly exposed area of downfaulted basalts and Middle Jurassic rocks on the lower slopes of Tom na Moine around Grid Sq. 475643. The rocks close to the contact, which is almost never exposed, are mostly faintly mottled and homogenous microgabbros. A small body of heterogenous rock is present at the eastern end of Druim na Gearr Leacainn ( 46786384 ). This is an augite - rich quartz diorite with strongly developed mottling: the composition of the rock varies between pyroxenite and diorite on a scale of several centimetres. Granular mafic inclusions are also present in this hybrid rock.

Although lacking large volumes of M1 felsic rocks, the microgabbros and country rocks between Dubh Chreag and Tom na Moine do contain extensive networks of felsic veins. These fall into two groups.

**Plate 3.47.** Coarse granophyric - textured granitic rock from vein cutting MBG microgabbros, north side of Druim na Gearr Leacainn. Note drusy cavity at left of field of view, partially filled with quartz and prehnite. Sample 65/2. Plane polarised light, field of view 3mm.



The older of the two is composed of irregular, unchilled but sharp - edged veins of relatively coarse, intensely altered aphyric granite with miarolitic cavities and well - developed coarse granophyric and/or poikilitic - quartz ( Lofgren 1971 ) textures ( Plate 3.47 ). The alteration of mafic minerals in these veins to granular clinopyroxene, hornblende and biotite, with subsequent partial retrogression to chlorite and epidote, suggests relatively high temperature hydrothermal metamorphism of these veins during the earlier part of M2 ( see 3.2.5.5 and Chapter 5 ). This implies that they were emplaced at or soon after the beginning of M2. The second group of felsic veins are fresher, fine - grained, joint - like and particularly intense in the vicinity of the pillowed hybrid granophyre/dolerite sheets also present in the area. These also cut Inner Series rocks to the north, on Beinn na Seilg ( see Map 1 and section 3.4.5 ), and appear to post - date the whole of the Hypersthene Gabbro.

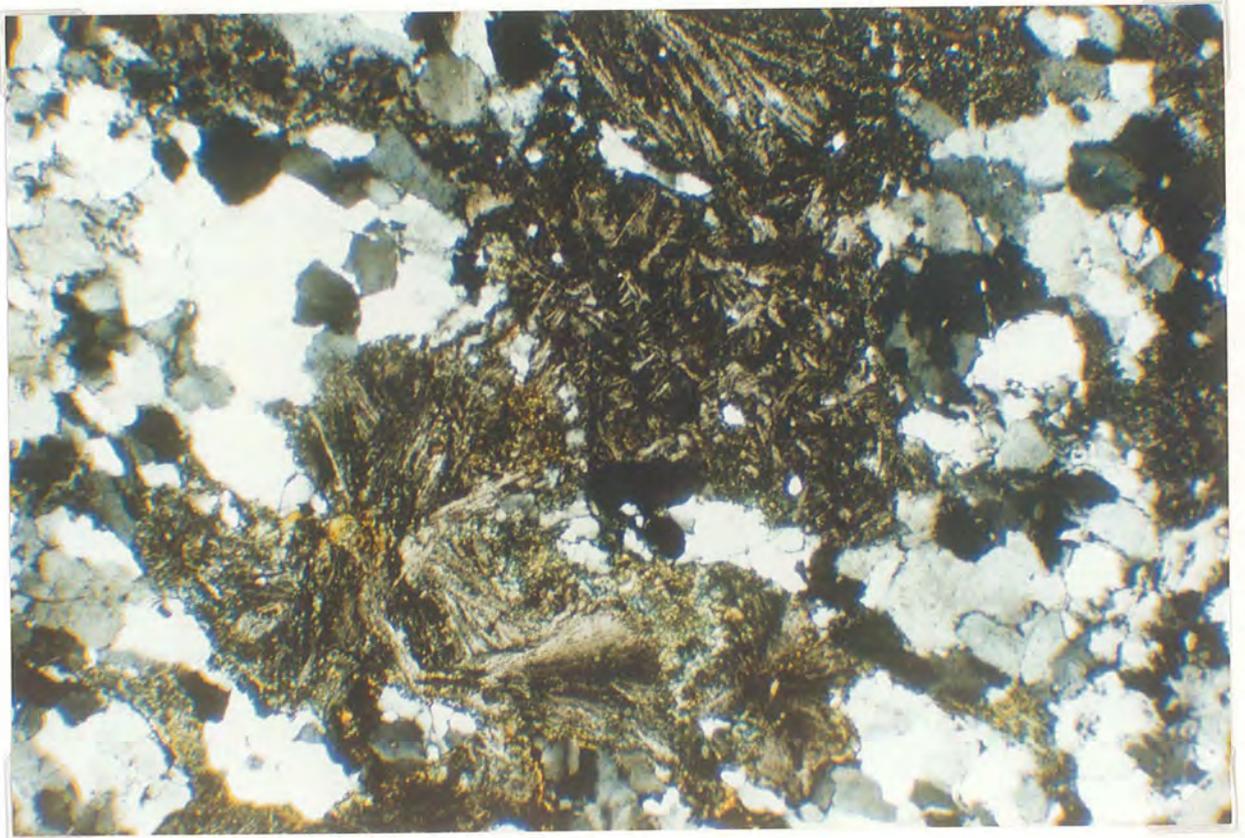
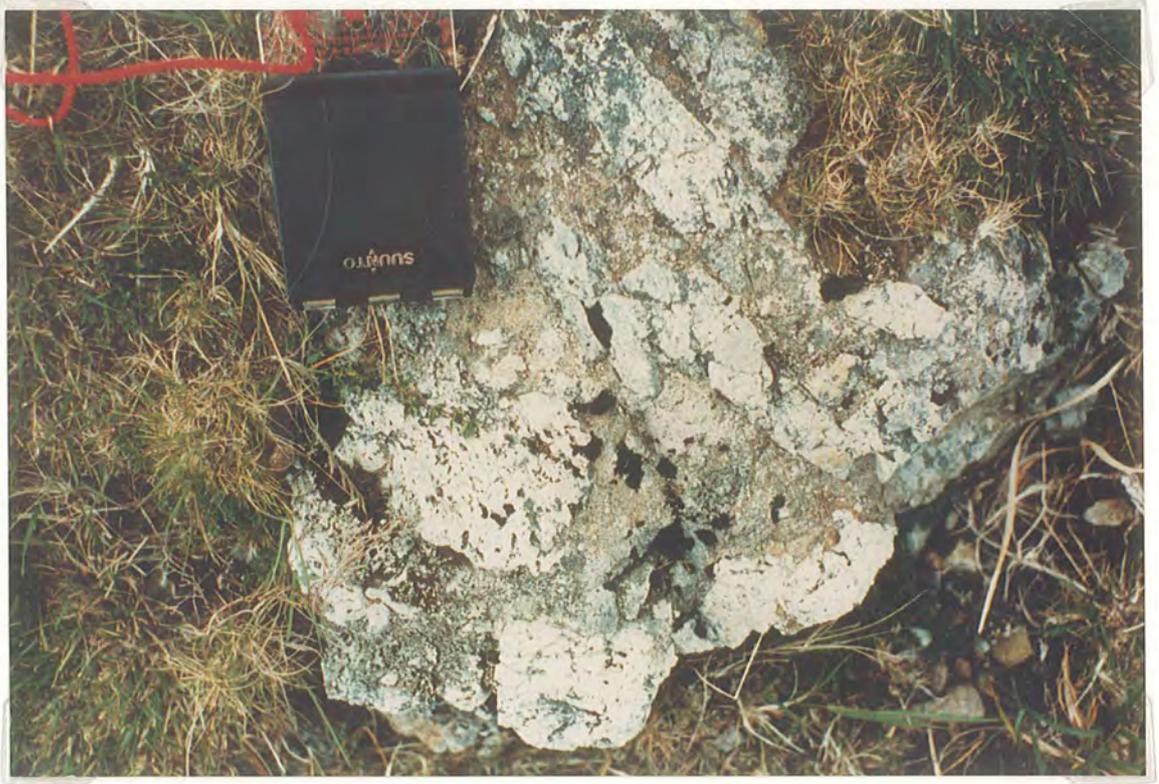
### **3.2.5.3. High degree partial melting of country rocks along the southern margin of the Marginal Border Group.**

The bulk of the wall rocks exposed along the southern margin of the pluton have moderately to highly refractory compositions. The high grade ( M1 ) metamorphism recorded in them ( section 3.2.5.5 ) has in general produced only small amounts of partial melting. However two, and possibly three, groups of rocks contain evidence for much higher degrees of partial melting and for large - scale melt mobility.

The first group is composed of small bodies of rheomorphic breccia at the contact on the eastern shore of the inlet at An Acairseid ( see above ) and at field location 240 on the south east side of Druim na Gearr Leacainn ( Grid Ref. 46656363 ). This latter occurrence is made up of two or more sheets intruded into pelitic hornfels in the transition zone between the Middle and Upper Pabba Beds. In contrast to the intensely recrystallised rocks at An Acairseid, these outcrops are relatively unaltered. Like the rheomorphic breccias at Glendrian Bay, about 20% of these rocks is made up of residual quartzite blocks. These range from compact, angular, coarse grained blocks with a small number of large melt pockets 3 to 4mm across, to irregular fine grained blocks with up to 40% pore space filled with devitrified glass ( Plate 3.48 ). The blocks and the pores are usually slightly elongate, both in the same direction. This suggests that the blocks are fragments of a bedded, slightly foliated protolith which was deformed either during the earlier stages of melting or in an

**Plate 3.48.** Rheomorphic breccia, field location 240. Note presence, within this one small outcrop, of a variety of residual block lithologies, ranging from solid quartzite to 'sponge' textured meta - sandstone with a high percentage of 'pore' space. Grid reference 46656363.

**Plate 3.49.** Sheaves and spherules of quench plagioclase within melt pockets in 'sponge' textured residual block in rheomorphic breccia, sample 240/2. Crossed polars, field of view 3.0mm.



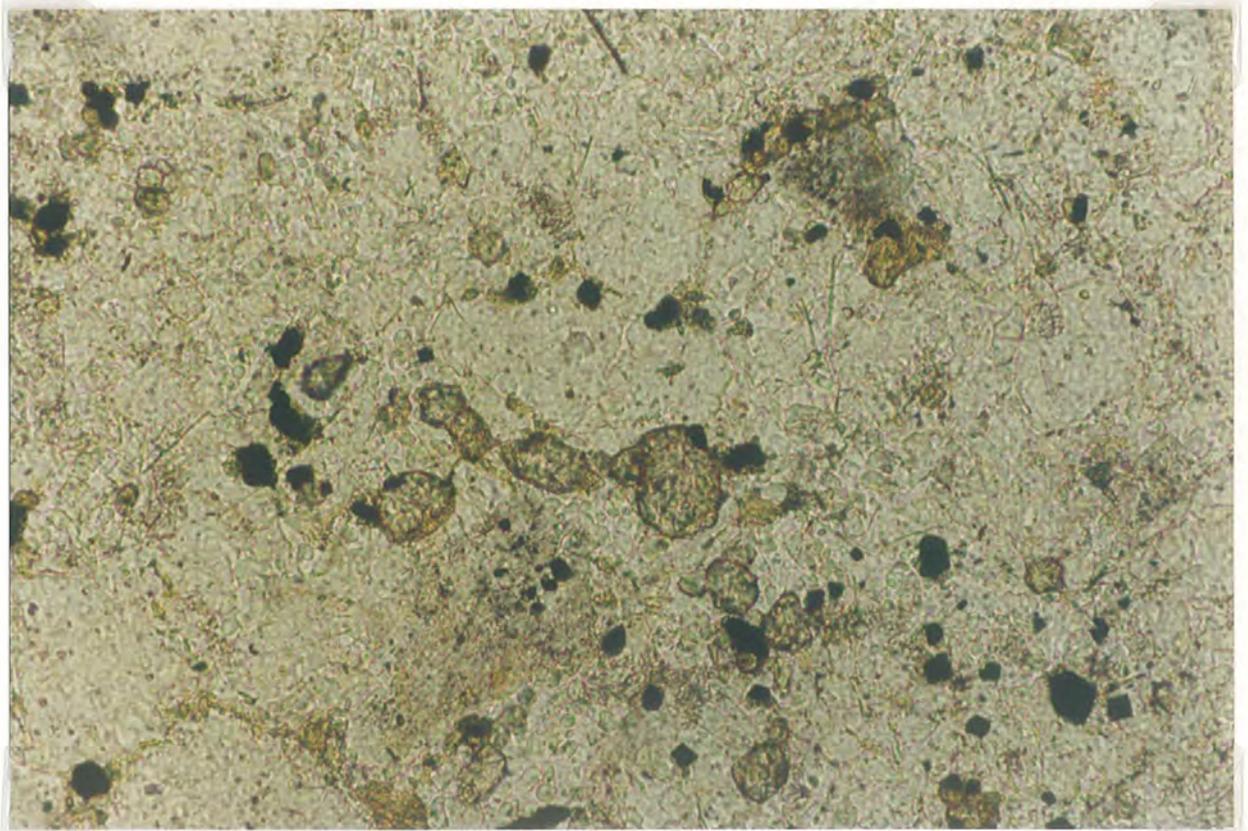
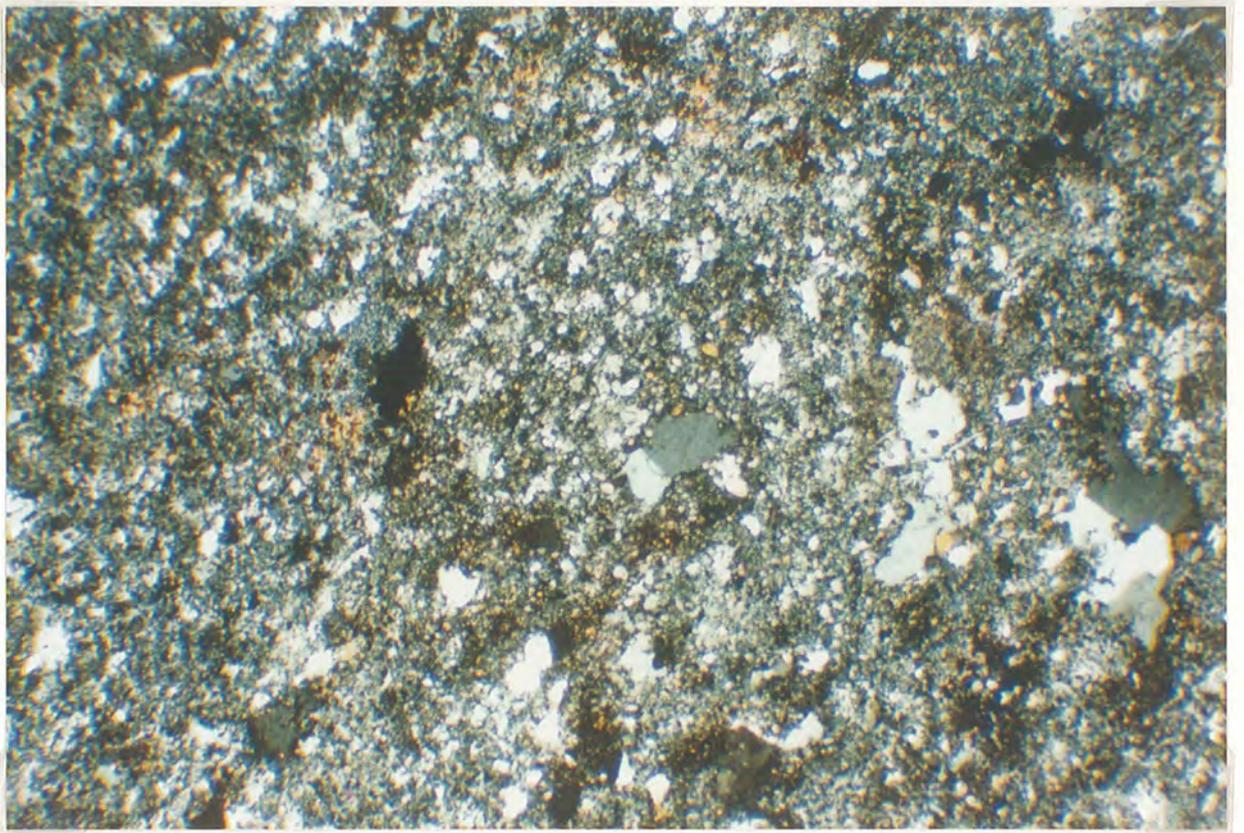
earlier event. These blocks differ from those in the Glendrian Bay area in being pure quartzite apart from the melt pores, and are mostly coarser grained as well ( Plate 3.49 ). The quenched melt in the pores and the matrix is also compositionally different, with a higher feldspar content and a higher content of mafic material: epidote and chlorite are the main alteration products and quench quartz is rare. The swirling clusters and sheaves of plagioclase in the quenched melt are indicative of very rapid cooling of a relatively Ca - rich melt ( Lofgren & Donaldson 1975 ). However the very rapid cooling of this rock cannot be used to infer that the host rocks, like those on the northern margin of the intrusion, were abruptly cooled at the end of M1, because its age relationships are not as clear as those of the rheomorphic breccias on the northern margin. All that can be said with certainty is that it was emplaced before the end of M2, because of the strong hydrothermal alteration present in the quenched melt.

The second group of rocks with evidence of high degrees of partial melting in this sector of the contact zone occur in the hanging wall block of the middle concentric fault on the south side of Hill 210, and also in the rocks immediately to the south of this fault, which are also in a downfaulted block ( see Figs. 3.10 & 3.13 ). These rocks are of pelitic and semi - pelitic composition: such rocks, as noted above, usually contain only small amounts of crystallised interstitial melt ( section 3.2.5.5, below ). In sections from these particular outcrops, however, the bulk of the rock is composed of a granophyric to granular - textured matrix, which contains corroded equant quartz grains, believed to be relict in origin ( as well as a small number of equant opaque grains with ragged margins which may also be relict ). The matrix contains abundant orthopyroxene, as well as cordierite or augite, in addition to the ubiquitous granophyric to poikiloblastic quartz - feldspar intergrowths ( Plates 3.50 and 3.51 ). The high degree of partial melting seen in thin section is also apparent in hand specimen and at outcrop in the form of a loss of bedding in all but the most refractory beds and residual blocks. Melt mobility is also apparent in the local development of rheomorphic breccia ( Fig. 3.12 ) and of diffuse veins cutting refractory lithologies, including metabasic hornfelses in the fault block around 44936326. The development of high degree partial melts during M1 in relatively refractory lithologies is another feature which is common to both the pseudoscreens on the northern margin of the pluton and the downfaulted blocks on the southern margin around Hill 210.

A third possible occurrence of high degree partial melts on the southern margin of the pluton is in the uppermost part of the Lower Pabba Beds in outcrops at the northern end of the Dubh Chreag

**Plate 3.50.** Largely fused pelite ( Middle Pabba Beds ) from downfaulted block south of Hill 210. Note presence of corroded residual quartz grains in granular to poikiloblastic quartz + feldspar + cordierite + orthopyroxene matrix. Sample 183D3. Crossed polars, field of view 3mm.

**Plate 3.51.** Subhedral orthopyroxene grains in interstices of quartzose semipelite hornfels. These appear to have grown from an abundant interstitial melt phase. Sample 183D2 ( Upper Pabba Beds ), from downfaulted block south of Hill 210. Plane polarised light, field of view 0.75mm.



gorge, just above the step in the contact at 45466328. Within about 20m of the contact, the bedding in the sediments becomes chaotic, and disappears altogether in exposures in the cliffs ( see Fig. 3.14, below ). Although it was not possible to sample these rocks, such loss of bedding and other primary structures is typical of rocks subjected to high degrees of partial melting, as at Loc. 183D and in many other examples of partial melting in both contact metamorphic ( for example, in Torridonian rocks close to the Rhum Complex ( Greenwood 1987 )) and regional migmatitic environments.

#### **3.2.5.4 The Structure of the Country Rocks away from the Contact in the An Acairseid – Kilchoan sector, and the Minor Intrusions which cut them.**

As much as 1.7km separates the outer contact of the MBG from the sea in this area. The rocks between the two were included in this study in order to investigate the implications of their thermal and structural evolution during the development of the Hypersthene Gabbro for the history of emplacement and heat loss of the pluton itself. This section deals primarily with the structures and minor intrusions observed in these rocks, whilst their metamorphic history is dealt with in section 3.2.5.5.

In addition to those structures which directly affect the geometry of the contact, dealt with in 3.2.5.2, three groups of structures may be distinguished which occur through out the sector. These are the outward tilting, of the Jurassic rocks in particular, which defines the dome first identified by Judd ( 1874 ); planar faults and dykes, which include both radial structures and others which are probably related to regional deformation; and the intense swarm of cone sheets, identified by Richey et al. ( 1930 ), which are associated with a number of concentric reverse faults. The two latter groups at least are superimposed on localised structures in the Maol Bhuidhe area ( Map 1 and Fig. 3.1 ) which are also included in this section.

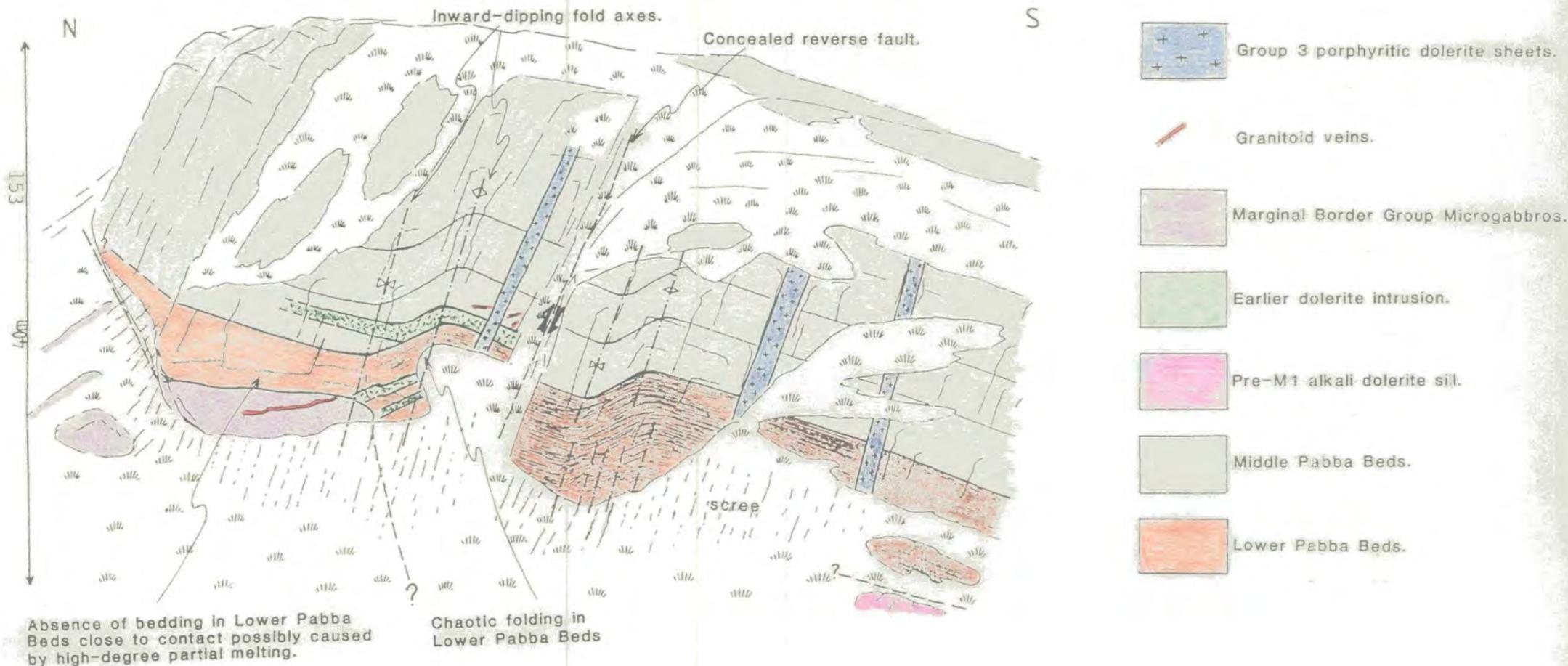
The dome structure around the Hypersthene Gabbro has in the past been considered to be concentric to the pluton and to 'Centre 2' of Richey et al. ( 1930 ). However, consideration of the outcrop of the sub - units of the Pabba Beds and the variation in strike directions in the Dubh Chreag to Tom na Moine area, covered by Fig. 3.11, suggests that this may only be approximately the case. Even allowing for the effect of radial faulting and the associated radial folding, the transition between the Middle and Upper Pabba Beds appears to rise towards the west at any given distance from

the contact. This seems to be an effect of the doming rather than a result of any increase in the thickness of cone sheets within the sequence from east to west, because these remain few and far between all along Druim na Gearr Leacainn. In addition, the strikes of the bedding in these rocks appears to be consistently more north - easterly than would be expected for a perfectly concentric structure. However, the difference between expected and observed strikes is only a matter of 5 to 10°, which may not be significant because of errors in measurement and possible distortions of the local magnetic field by the adjacent gabbros, which would affect the compass used in strike measurements. There is nevertheless a possibility that the centre of the dome is displaced westward from that of the Hypersthene Gabbro, or that the dome is elongated north - south relative to the pluton, but this is by no means proven by the existing observations. Exposures of domed rocks elsewhere around the contact are either too limited, or too strongly affected by later deformation, to distinguish between these two possibilities.

Three concentric regions can be distinguished within the dome structure on the basis of changes in dip and in minor structures. Close to the contact, in the downfaulted blocks on Dubh Chreag, on Druim na Gearr Leacainn and east of Tom na Moine, dips in the Jurassic rocks vary but are often in excess of 50° outwards, especially in the block east of Tom na Moine, where dips are vertical in places. These steep dips could in part reflect rotation of the sediments during movement on the concentric faults but there is no other evidence to suggest that the faults have a listric geometry. If they did, this would require those shallowly - dipping rocks also found within the fault blocks, at 45966342 and 45656344, for example, to have previously had inward dips of as much as 30°. Although such inward dips are found in the Glendrian Bay area ( 3.2.2 ), they are not found anywhere on the southern margin of the intrusion, except in small monoclinial folds such as those exposed on the eastern side of the Dubh Chreag gorge ( Fig. 3.14 ). Overall, this innermost exposed zone of the dome appears to have somewhat variable but largely outward dips, and to contain a number of small inward - verging folds.

A second region of the dome extends from within about 100m of the contact to 700 to 1000m from it. Except close to the radial faults and folds, this is characterised by relatively uniform outward dips, possibly increasing outward on average, in the range 25 to 35°. Minor fold structures are rare. The outer limit of this zone is only exposed south of Dubh Chreag, where it appears to be a gradual transition to the more shallowly - dipping third zone.

Fig. 3.14. Sketch of the eastern side of the Dubh Chreag gorge at the outer contact of the Marginal Border Group.



The main exposures of the third region of the dome are in outcrops of the Upper Pabba Beds along the coast of Kilchoan Bay, between 471627 and 481635. Dips in these vary from  $10^{\circ}$  to as much as  $70^{\circ}$  outwards, although dips steeper than  $30^{\circ}$  are rare. The strike of the bedding also varies, from 050 to 060 at the southwest end, around 471627, to about 080 around 475633, to about 050 around 481637, quite apart from minor folding in the vicinity of radial faults. The overall outward dips have been modified in this third zone by a series of open outward - verging concentric folds. The folds die out laterally over distances of a few hundred metres: this is best seen in the case of the tightest fold, which runs along the shore near high tide for only 200m near 475633. They affect both the bedding and a weak micaceous and stylolitic foliation which is present in pelitic horizons.

Several constraints can be placed on the age of the dome structure. It **must** predate the concentric inward - dipping faults as otherwise these would have originally had an orientation appropriate to ring dyke fractures ( Anderson 1936 ) and would have ring dyke intrusions emplaced up them. With only one possible exception, all the cone sheets in the area cross - cut the minor folds associated with the doming: in contrast, the minor folds do affect the foliation in the pelites and must post - date it. This fabric is shown in section 3.2.5.5 to have formed in a phase of metamorphism, M0, prior to the emplacement of the Marginal Border Group. Although there are problems in correlating episodes of intrusion close to and far from the Hypersthene Gabbro ( see below ), these age relationships, particularly between the cone sheets and the folding, do suggest that the doming predated the emplacement of the Marginal Border Group in its present position. The same age constraints apply to open folds, with near - radial fold axes, associated with the broadly radial north - south to NNW - SSE, steeply dipping faults which also cut the area. In contrast, the faults themselves show evidence for activity at a wide variety of times.

In the western part of the area these faults, and the dykes which are intruded up some of the fault planes, fall into two distinct groups: a north - south set, truly radial to the contact, some at least of which have acted as compartmental faults to the concentric inward - downthrowing fault system as described in section 3.2.5.1, and a set which strikes  $140^{\circ}$  to  $155^{\circ}$ . Further east, any faults which did develop radially to the contact as a result of magmatic stresses would in any case strike roughly parallel to the second group. This second group runs parallel to the main dyke swarms in northwest Scotland and may be related to regional rather than local deformation ( see Chapter 7 ).

The radial faults close to the contact have been dealt with in section 3.2.5.1. Elsewhere, numerous faults are present along the coast between Maol Bhuidhe and Kilchoan. Most of these cannot be traced inland because of poor exposure. However, some are in line with prominent radial gullies on Druim na Gearr Leacainn which conceal faults that offset cone sheets in their walls. Most are subvertical, although several high - angle extensional faults and a single reverse fault are present, all with dips of  $80^\circ$  or more. Movements on all of these faults, except for a few large ones such as the Dubh Chreag gorge fault and others marked on Fig. 3.10 and Fig. 3.11, are only 10m or less. The sense of movement on the faults varies widely. The early folding associated with the faults records dip - slip movement, whilst later dip - slip motion is implied by opposite offsets of outward - dipping sediments and inward - dipping cone sheets, for example at 47866347 and 47896349. Apart from early faults in the Maol Bhuidhe area and those close to the contact, dip - slip movements were identified on 8 faults, of which 6 had easterly downthrow. This asymmetry, which suggests uplift to the west of the Kilchoan shore section, is consistent with the evidence for non - concentric doming noted above. In the case of the later faults which displace the contact of the MBG or cut cone sheets, later strike - slip motion is more common than dip - slip. Both sinistral and dextral movements were identified, although the larger strike - slip faults are all dextral.

Radial and/or regional dykes are much more common than the corresponding faults. Where they are intruded along fault planes they are usually post - deformational. There may, however, be a sampling bias at work, because the post - dyke deformation usually takes the form of hydrothermal veins with lateral offsets of the vein walls. Exposed examples include a dyke at 47896349, where calcite veins form a set of en echelon fractures, connected by shear fractures, with dextral offset, and the Dubh Chreag gorge fault, which contains a deformed and intensely altered dolerite dyke. More commonly, deformed intrusions are very badly exposed because of preferential erosion of the altered rock. Consequently, deformed radial intrusions tend to be lost in the floors of the radial gullies which mark many of the faults. The dykes vary widely in composition, from alkali gabbros and dolerites through feldspar - phyric dolerites to composite acid - basic dykes and a number of pitchstones and felsites. They are also very varied in their age relationships. Some alkali dolerites and felsites predate the early M0 metamorphism ( defined in section 3.2.5.5 ), for example dykes of the regional, NNW - SSE striking, group on the southwestern side of Maol Bhuidhe. Others are very late, cutting the Inner Series of the Hypersthene Gabbro and later Centre 2 intrusions as well. However, in only one case can a dyke be shown to cut an earlier cone sheet and be cut in turn by

a later one. This is at 46976278: the cone sheets concerned belong to Groups 2 and 3, as defined below, respectively. Possible reasons for this lack of dyke emplacement events during the period(s) of cone sheet emplacement are discussed in Chapter 7.

The cone sheets in the area south of the Hypersthene Gabbro were regarded by Richey et al. ( 1930 ) as all belonging to an outer set of cone sheets confocal to their Centre 2 of the Ardnamurchan central complex, and to all originate from a common focus a few kilometres below the surface. Investigation of the cone sheets during the course of the present work indicates that they are much less regular, and considerably more complex, than this generalisation implies. Although it was not possible to map all the cone sheets individually in the course of this study, and in general only the large and distinctive sheets of the Group 3 defined below were ever mapped individually, the following points were noted in the field:

1). As is the case on the northern margin of the pluton, most cone sheets occur in swarms. Areas up to 1km long concentric to the Hypersthene Gabbro, and a few hundred metres wide at most, are almost entirely formed by cone sheets, of Groups 1 and 2 ( defined below ) in particular. Outside the swarms, cone sheets form less than 25% of the outcrop, and usually less than 10% . Group 3 cone sheets do not show the phenomenon to the same degree: they are sparsely distributed throughout most of the area, although they do form as much as 30% of the outcrop south of Dubh Chreag, around 451631.

2). Individual sheets can be traced for tens to hundreds of metres, particularly in cliff exposures and also in the case of the distinctive, outweathering coarse dolerite sheets of Group 3 ( defined below ). One member of this group can be traced laterally for almost 1km along the south side of Druim na Gearr Leacainn and shows little thickness variation in this distance.

3). Most of the sheets dip concentrically inwards, although at a wide variety of angles even within individual localities, are planar or slightly curved overall, but show numerous small - scale irregularities. Although few cone sheets can be traced for a sufficient distance laterally for a curved or conical geometry to be apparent, this is invariably present in the few sheets which can be traced for more than 0.5km or so. Individual sheets show a wide variety of departures from the simple confocal, inverted conical shape originally proposed for this class of intrusion.

One of the commonest, which may contribute to the difficulties of tracing individual sheets between outcrops, is the development of ramp - like structures on which the cone sheets cut up or down section along strike ( Fig. 3.15A ). Larger - scale versions of these structures may account for a number of inclined sheet intrusions which are not concentric to the pluton and strike obliquely to adjacent cone sheets, at bearings closer to north - south. These, however, occur in groups with distinct ages relative to cone sheets at the same localities: for example, in 470627, fine - grained dolerite sheets with strike  $020^{\circ}$ , four dipping  $30^{\circ}$  northwest and two dipping shallowly southeast, cross - cut all the cone sheets in the outcrop. An alternative interpretation of these inclined intrusions is presented in Chapter 7.

As on the northern margin of the Hypersthene Gabbro, a number of outward - dipping dolerite sheets, with dips in the range  $20 - 40^{\circ}$ , are associated with the cone sheets. Although a number of these are more or less parallel to bedding, particularly along the coast at Kilchoan, most are not, and cross - cut the bedding at angles of several degrees or so. In a number of cases, outward - dipping sheets pass up- or down - dip into normal cone sheets, for example that shown in Fig. 3.15B. The intersection between the two halves of the intrusion has a distinct eastward plunge, indicating that one or both may not be confocal to the pluton, although this may be the result of later tilting rather than a primary feature. It is not known how many cone sheets show this, because the necessary 3-D exposures are lacking.

In contrast to the outward - dipping dolerite intrusions on the northern margin of the Hypersthene Gabbro there is little evidence to suggest that these outward - dipping sheets on the southern margin are associated with thrust faulting. Where an intrusion has inward - and outward - dipping segments exposed, they are of similar thickness. This implies that there was little relative horizontal movement of the top and bottom surfaces: in other words, that the direction of opening or dilation of the intrusion was near - vertical. In a large number of cases, it is possible to determine the dilation directions of cone sheets from the offsets of markers, such as the intersections of distinctive beds with the sheet walls, or matching irregularities in the walls of the intrusions themselves, and orthogonal to vertical dilations are typical ( see Chapter 7 for further discussion of cone sheet emplacement mechanisms and their relationship to the stress field created by the Hypersthene Gabbro ).

A small number of irregular inward - dipping dolerite sheets are, however, associated with members

FIG. 3.15A CONE SHEETS WITH OBLIQUELY DIPPING SEGMENTS, IN SOUTH - FACING CLIFFS AT SRON BHEAG.

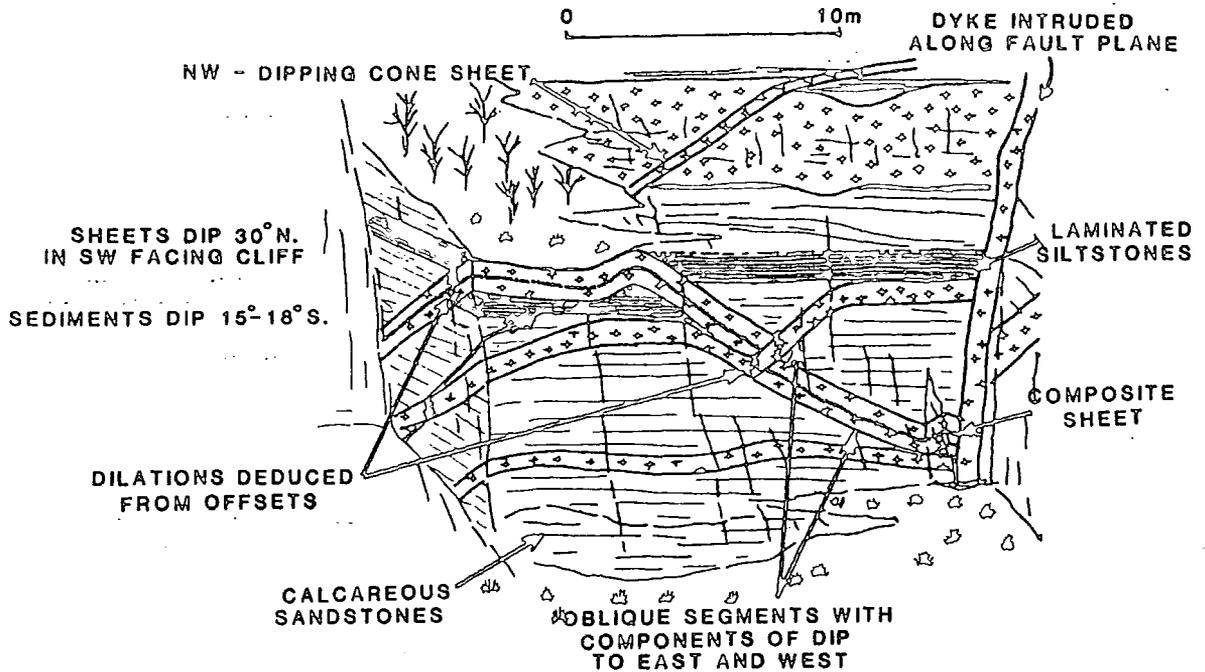
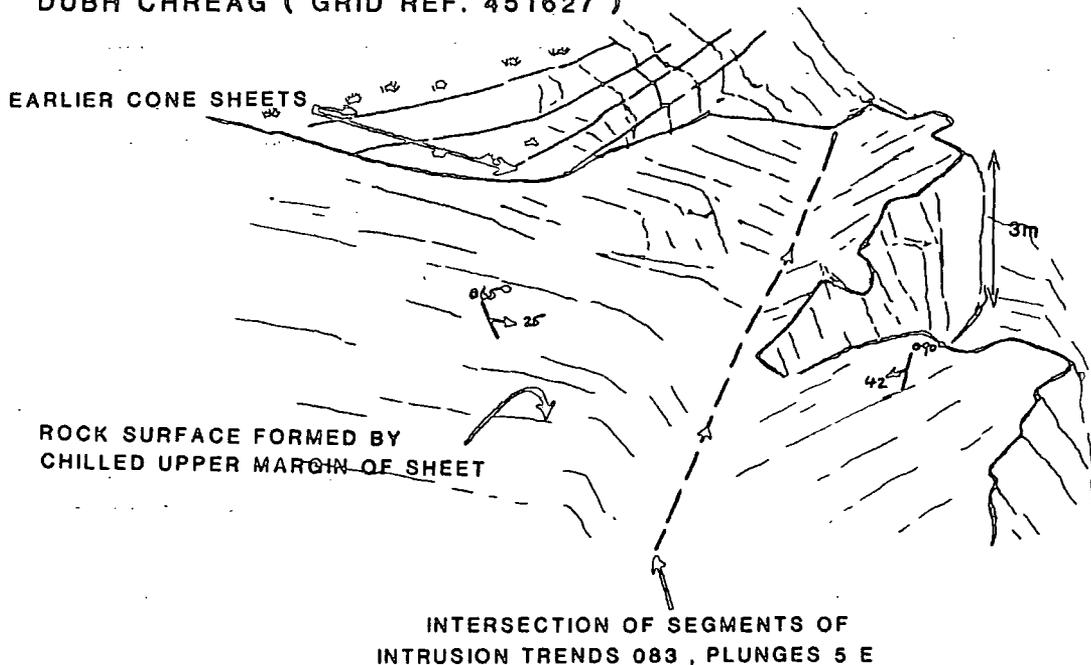
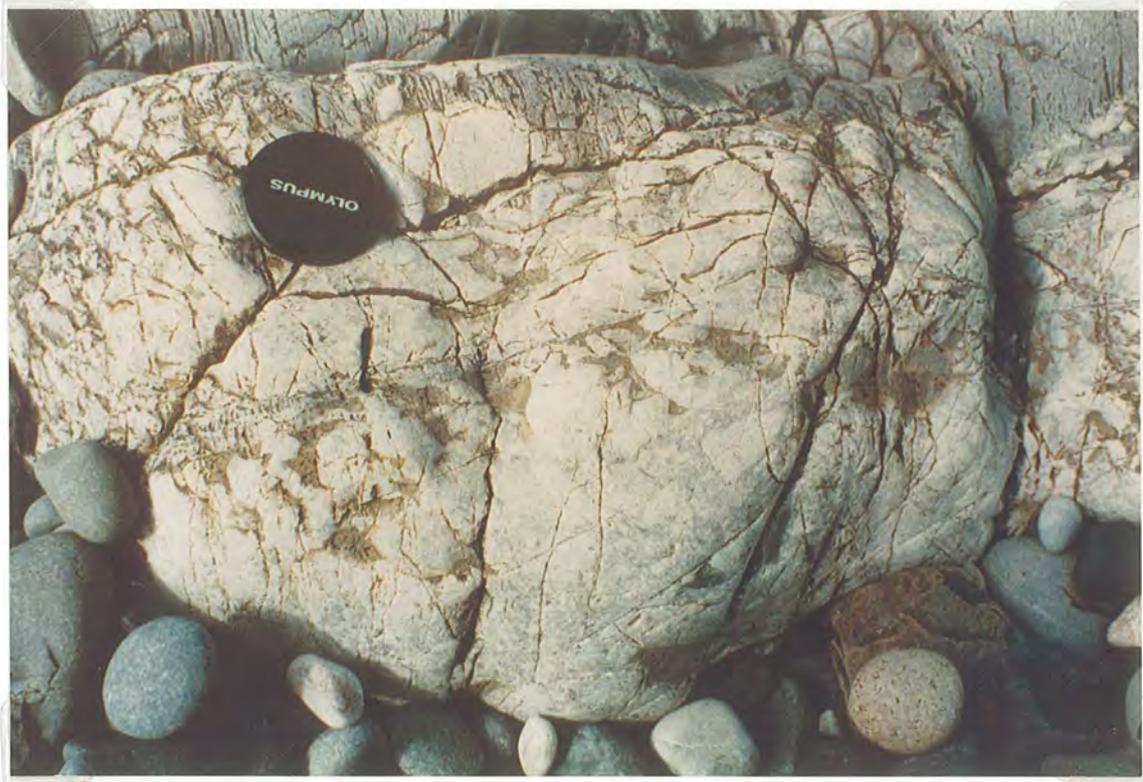


FIG. 3.15B. LINE DRAWING FROM PHOTOGRAPH OF DOLERITE CONE SHEET PASSING NORTH INTO SOUTH - DIPPING DOLERITE SHEET, SOUTH OF DUBH CHREAG ( GRID REF. 451627 )



**Plate 3.52.** Brecciated rock at base of dolerite cone sheet, field location 344 ( grid reference 45506277 ). The interstices between the clasts in the fault breccia are filled with chilled basaltic material close to the contact with the dolerite. This implies that the breccia was invaded by magma soon after it formed ( see text ).

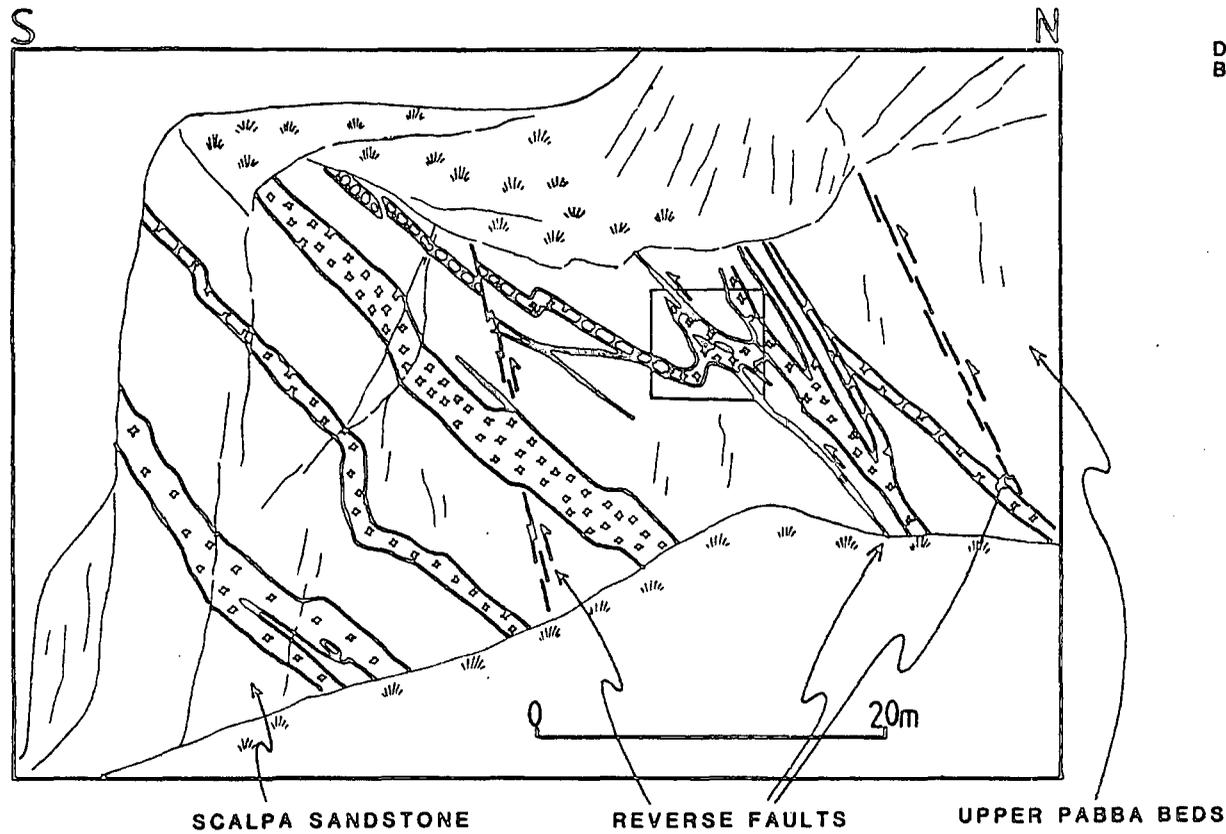


of a group of steeply inward - dipping concentric reverse faults whose ages can be shown to overlap with those of the cone sheets. The best examples of this relationship are exposed in cliffs at the lower end of the Dubh Chreag gorge, at about 45376280 ( Fig. 3.16 ), and on the western coast of Maol Bhuidhe at 46036233. In these outcrops, irregular dolerite sheets are apparently offset by the faults but the fault planes themselves are intruded by veins of dolerite rooted in the main sheet intrusions. This implies simultaneous or overlapping faulting and thrusting. Similarly, the poorly exposed base of a thick ( 10m+ ) dolerite cone sheet at 45506277 has veins of chilled basaltic material extending from it into the interstices of an underlying fault breccia and taking the place of the epidotic and chloritic material which fills the voids between the breccia clasts further from the contact with the dolerite ( Plate 3.52 ). This implies that intrusion of the dolerite immediately postdated formation of the fault breccia, which is consistent with the presence of minor steeply north - dipping faults which cut both breccia clasts and veins. Other high - angle reverse faults occur in the Maol Bhuidhe area ( see below ), at the northern end of the Dubh Chreag Gorge at about 454634 ( see Fig. 3.14 ) and along the Kilchoan shoreline around 475633. All except the poorly exposed fault at the north end of the gorge can be constrained, chiefly by cross - cutting relationships with the cone sheets, to have formed during the M2 phase of metamorphism defined below ( section 3.2.5.5 ). In contrast, low - angle thrust structures like those on the northern margin of the pluton are absent in the An Acairseid - Kilchoan sector, apart from a few tiny centimetre - scale inward and outward - dipping thrusts close to the contact and a single subhorizontal to very shallowly outward - dipping shear zone exposed over a distance of several metres in a radial gully on the western side of Tom na Moine ( Grid Ref. 46846377 ). Veins developed at extensional jogs within this shear zone contain the assemblage epidote + alkali feldspar, indicating that it is of late M2 age ( see section 3.2.5.5, below ) as earlier M2 metamorphism on Tom na Moine is of higher grade.

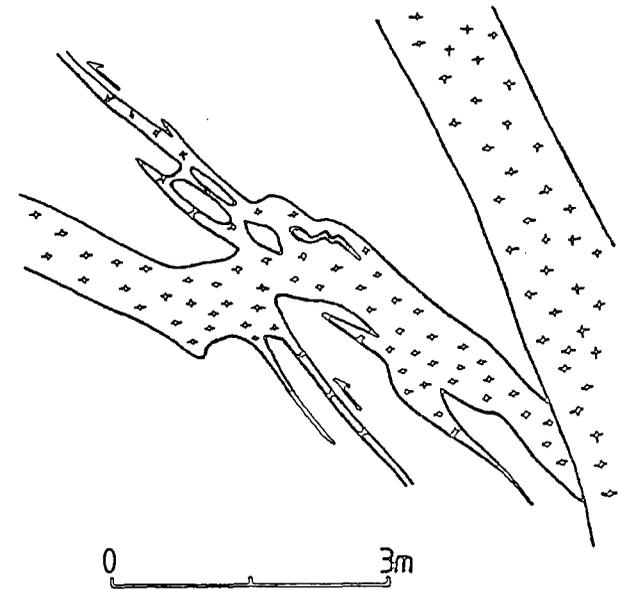
Determination of the age relationships of the cone sheets to the Marginal Border Group in this area is difficult because of the generally poor exposure of the contact zone, except near An Acairseid and south of Hill 210, and also because of the limited extent and poor exposure of the contact zone itself. It is however possible to distinguish four groups within the cone sheets and related intrusions in the area, on the basis of consistent cross - cutting relationships between members of the different groups, original igneous petrography, and the different phases of metamorphism exhibited by them ( see section 3.2.5.5 ).

FIG. 3.16. LINE DRAWING, BASED ON FIELD PHOTOGRAPHS AND SKETCHES, SHOWING GROUP 2 CONE SHEETS AND COEVAL HIGH - ANGLE REVERSE FAULTING ON THE WEST WALL OF THE DUBH CHREAG GORGE AT ITS SEAWARD END.

161



DETAIL OF CONE SHEET OFFSET BY, BUT VEINING, REVERSE FAULT.



Group 1 cone sheets include the granular textured, aphyric sheets in the An Acairseid area and on Hill 210 amongst others. A feature of some of these sheets in the An Acairseid area, found in no other minor intrusions in this area except some of the much later Group 4, is that they contain mixed acid and basic rocks. The felsic component occurs as diffuse coarse - grained lobate inclusions in otherwise mafic intrusions ( e.g. at 43846315 ) or is the dominant component of heterogeneous sheets with fine - grained mafic and hornfelsed quartzose inclusions: the best example of the latter is an early sheet on the coast at An Acairseid ( 43616301 ) which is cut by basic Group 1 sheets in the area. Group 1 sheets consistently dip inwards, at angles varying from 20° to near - vertical. They are the only group to definitely be truncated by the contact zone of the MBG. Close to the pluton they appear to have been affected by M1 but outside the M1 aureole it is in general not possible to directly distinguish Group 1 intrusions from Group 2.

Group 2 includes the vast majority of the cone sheets in the An Acairseid to Kilchoan sector. They are sparsely feldspar - phyric, mainly rather fine - grained doleritic rocks, with a very dark colour in hand specimen caused by schillerisation and clouding of the plagioclase. Secondary pyrite films and flakes are another characteristic feature used for field identification of this group. Primary ophitic to sub - variolitic textures are present, although the rocks have been affected by M2 metamorphism. Feldspar glomerocrysts and sub - granular feldspathic microxenoliths are present in some samples of this group. The larger feldspar phenocrysts have distinct rounded unzoned cores, but these are much less abundant than in Group 3 dolerites ( see below ). Petrographically, the Group 2 dolerites are very similar to the sparsely porphyritic M2 cone sheet dolerites found on the northern margin of the intrusion ( section 3.2.2 ). Group 2 includes all the irregular and outward - dipping sheet intrusions observed in the field. Members of the group are usually truncated by movement on the radial faults and may also be truncated by later movements on the concentric inward - dipping faults, for example on Druim na Gearr Leacainn.

Group 3 are large, relatively regular, sparsely distributed cone sheets which mostly occur within 1km of the contact. They are very distinctive in the field, being pale, coarse - grained dolerites with abundant large plagioclase porphyrocrysts. These are so abundant in a few sheets, such as that which cross - cuts the Druim na Gearr Leacainn fault system at field location 240, that the sheets concerned are near - anorthositic. All dip inwards, at angles ranging from 20° to more than 70°. They commonly postdate most movement on the radial faults, although they are usually offset by

the NNW - SSE trending faults, and invariably post - date all significant movement on the concentric normal faults. In addition, they are only affected by the later, low grade part of the M2 phase of metamorphism defined below.

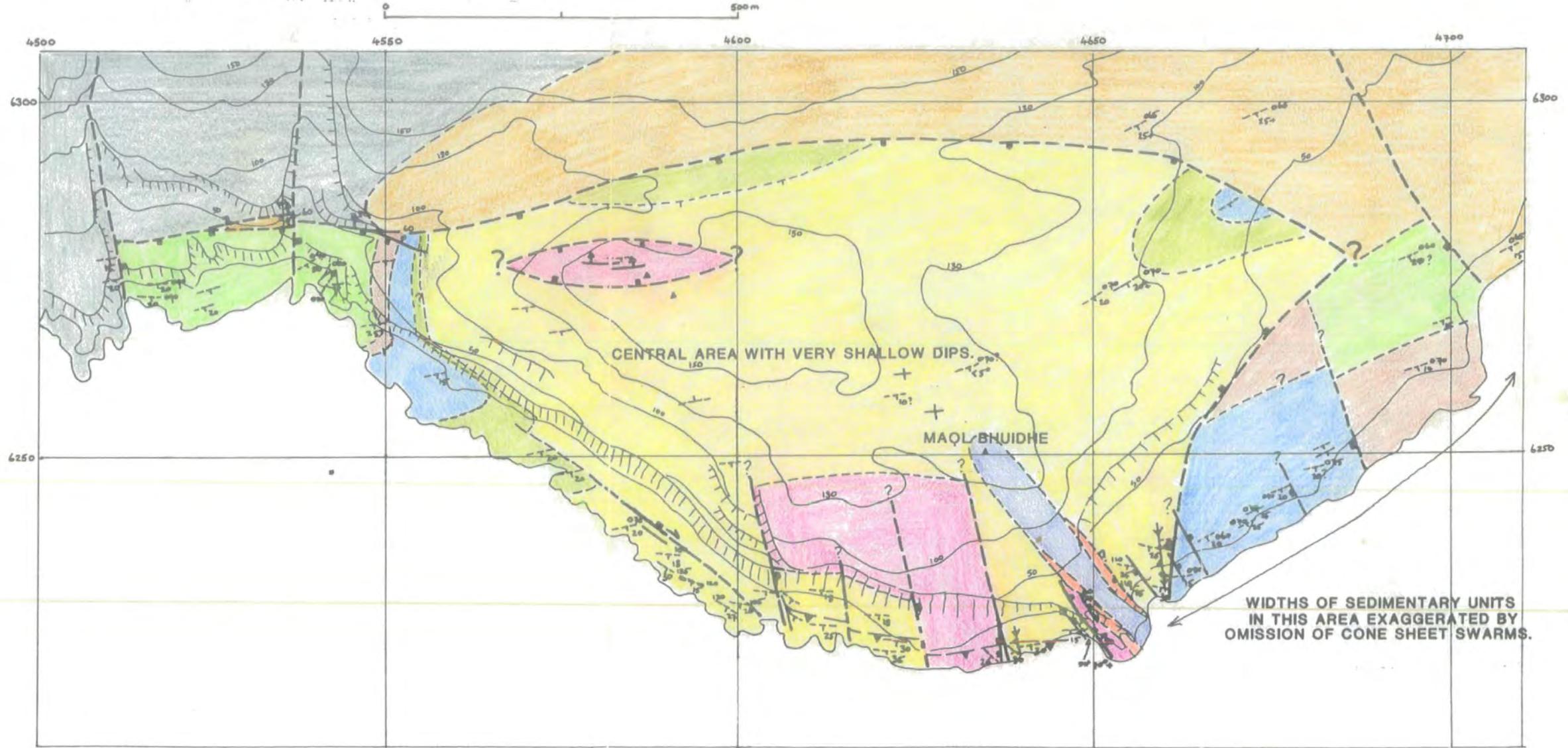
Group 4 is defined purely on age relationships, being very varied in composition. It consists of a variety of late intrusions which cross - cut both MBG and Inner Series rocks at An Acairseid, and similar small intrusions elsewhere. They include porphyritic dolerites with prominent chilled margins and a strong zonation from aphyric margins to a porphyritic centre, pillowed hybrid intrusions ( with strongly chilled basaltic pillows in a microgranophyric host ) and porphyritic and aphyric microgranophyres. All show only moderate degrees of chloritic and epidotic alteration. Most occur in planar sheet intrusions. These are mainly steeply inward- or outward - dipping and have a limited range of strikes (  $070^{\circ}$  to  $100^{\circ}$  ), implying that they are not concentric with the pluton: they are certainly oblique to many of the earlier cone sheets.

The structures on and around the hill of Maol Bhuidhe ( Map 1 ) are dealt with separately in this section because in large part they predate those described above and are not simply concentric or radial to the Hypersthene Gabbro. Broadly speaking, Maol Bhuidhe is composed of a block of Middle Jurassic sediments and Tertiary basalts which has been downfaulted, at least on its northern and southeastern sides, against Liassic sediments ( Fig. 3.17 ). The most distinctive structural feature of the area is that much of it is subhorizontal to gently dipping, rather than forming part of the dome structure, concentric to the margin of the Hypersthene Gabbro, which dominates the rest of the area to the south of the pluton.

The structures in Fig. 3.17 fall into three groups: (1) North - south and NNW - SSE faults, and associated folds, which are particularly prominent at the margins of downfaulted blocks of Tertiary volcanic rocks within the Maol Bhuidhe block; (2) the bounding fault on the southeastern margin and an analogous but much smaller fault on the southwestern coast; (3) the long - lived fault on the northern side of the Maol Bhuidhe block.

The north - south trending subvertical faults which bound the downfaulted block of basalts at Sron Bheag are associated with abundant wairakite (?) + calcite and albite + epidote + actinolite veins running parallel to the fault planes. They were, therefore, active during the pre - Hypersthene

FIG.3.17. STRUCTURE OF THE AREA AROUND MAOL BHUIDHE.  
 CONE SHEETS, DYKES AND MINOR FELSITE INTRUSIONS OMITTED FOR CLARITY.  
 TOPOGRAPHIC CONTOURS IN m.



- Microgabbro.
  - Felsite.
  - Tertiary basalts.
  - Middle Jurassic:
  - Calc-silicate sandstone.
  - Meta-siltstones.
  - Calc-silicate meta-limestones.
- } Cuingleum vent intrusion.

- Upper Lias: pelite with meta-ironstone.
- Middle Lias: calc-silicate meta-sandstone.
- Lower Lias:
- Upper Pabba Beds
- Middle Pabba Beds.
- Monoclinial Fold.

- Plunge direction of fold axis
- Fault
- Strike-slip fault.
- Bedding dip & strike.
- Horizontal beds.

Gabbro phase of hydrothermal metamorphism M0 defined below, in section 3.2.5.5. Both the faults and the veins are cross - cut by aphyric dolerite cone sheets of Groups 1 or 2. In common with other early radial faults, the eastern fault at Sron Bheag has a prominent monocline associated with it, which has the same sense of displacement as the fault. Similar but larger monoclines are present on both sides of the composite intrusion and vent at Cuingleum ( Grid Sqs. 464623 and 465623 ). As noted in Chapter 2, slumping of the wall rocks into this vent occurred during its formation: the development of the adjacent monoclines may also be related to the vent rather than to regional deformation. The vent and associated structures are all cross - cut by numerous Group 1 and/or Group 2 cone sheets.

The faults on the southeastern and southwestern sides of Maol Bhuidhe are also cut by cone sheets and are associated with M0 hydrothermal veining. Both faults downthrow towards the centre of the Maol Bhuidhe block but cut rocks which dip steeply away from the block, as do the rocks along the southern shore between the two faults ( Fig. 3.17 ). The structures around these faults thus record a period of relative uplift of the Maol Bhuidhe block, whilst the subsequent faulting produced relative subsidence. The M0 age of this activity indicates that it would have taken place during or before the doming prior to the emplacement of the Hypersthene Gabbro but the change in the sense of motion suggests that the relationship between the two is not simple, and that additional causes must be sought for the deformation in the Maol Bhuidhe area ( see Chapter 7 ).

The northern margin of the Maol Bhuidhe fault block is formed by a near - vertical fault, now concealed except at its western end, around 454627. At this point it dips north at more than 80° and is marked by a zone of hornfelsed veined and brecciated sediments. To the south, around 45486283, the Middle Jurassic rocks are cut by a north - dipping ( at 60° ) reverse fault which is itself truncated by the main fault. Reverse faults dipping north at 50 to 60° are also exposed in the wall of the Dubh Chreag gorge to the west ( see Fig. 3.16 ). However, these place Middle and Upper Pabba beds above Scalpa Sandstone, implying transfer, along the Dubh Chreag gorge fault, of some of the offset on the main fault ( which places the same rocks above Middle Jurassic rocks ) to faults offshore to the south. The cone sheets in Fig. 3.16 belong to Groups 1 and/or 2, whilst the fault on the north side of Maol Bhuidhe offsets Group 1 and/or 2 sheets but not a Group 3 sheet at about 460629. Age relationships between the cone sheets and metamorphism ( section 3.2.5.5 ) therefore imply an M1 or early M2 age for the smaller reverse faults and cessation of activity on

the larger fault before the latter part of M2, which is consistent with the hornfelsed character of its fault breccia.

The main fault, however, is more steeply inclined than the contemporaneously - active smaller concentric reverse faults. In addition it appears to have accommodated the movement on the fault on the southeastern side of Maol Bhuidhe which occurred during M0 ( see Fig. 3.17 ). To do this it must have been in existence during M0 and have dipped south rather than north. This can be explained by supposing that this early movement took place prior to the doming, which is consistent with the age relations of the faults, the doming and M0 (see section 3.2.5.5 ). Subsequent rotation of the faults through about 30° during the doming would turn the northern fault into a reverse fault which could take up some of the central uplift implied by the adjacent reverse faults. This model supports the suggestion made above that the faulting in the Maol Bhuidhe area is not simply related to the doming, as implied by certain recent models of deformation around the Tertiary Igneous complexes in general ( R.W.England pers. comm. ).

#### **3.2.5.5. Metamorphism on the southern side of the Hypersthene Gabbro and its Age Relationships to Deformation and the Emplacement of Minor Intrusions.**

Three phases of metamorphism can be recognised along the southern margin of the Hypersthene Gabbro. The relatively poor exposure of the contact and the lack of continuous exposure further away from it forces reliance on these episodes of metamorphism as means of correlating and dating intrusions and structures rarely actually seen to cross - cut each other. The validity of this procedure is critically dependent upon the extent to which the episodes of metamorphism, like the episodes of emplacement of groups of minor intrusions, can be treated as synchronous over wide areas, and the extent to which departures from synchronicity can be recognised and allowed for. This problem is discussed further in section 3.2.5.6. and in Chapter 5.

The first phase of metamorphism to the south of the pluton is here referred to as M0, because it predates the emplacement of the intrusion in its present position and may not be directly related to it. In contrast to the situation on the northern margin of the pluton, where it is all but obliterated by the effects of M1 metamorphism in that area, M0 is well developed on the southern margin, both inside and outside the later M1 aureole. The effects of M0 on different lithologies are here considered

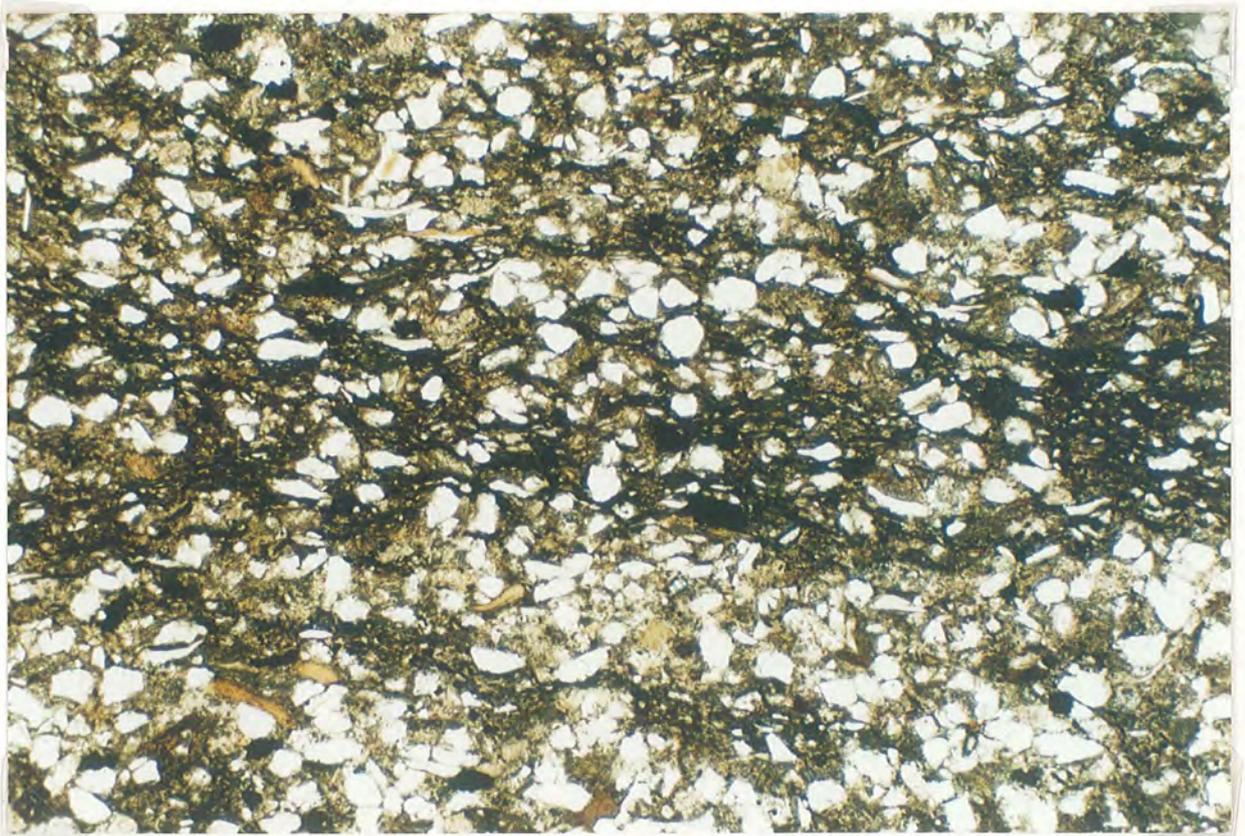
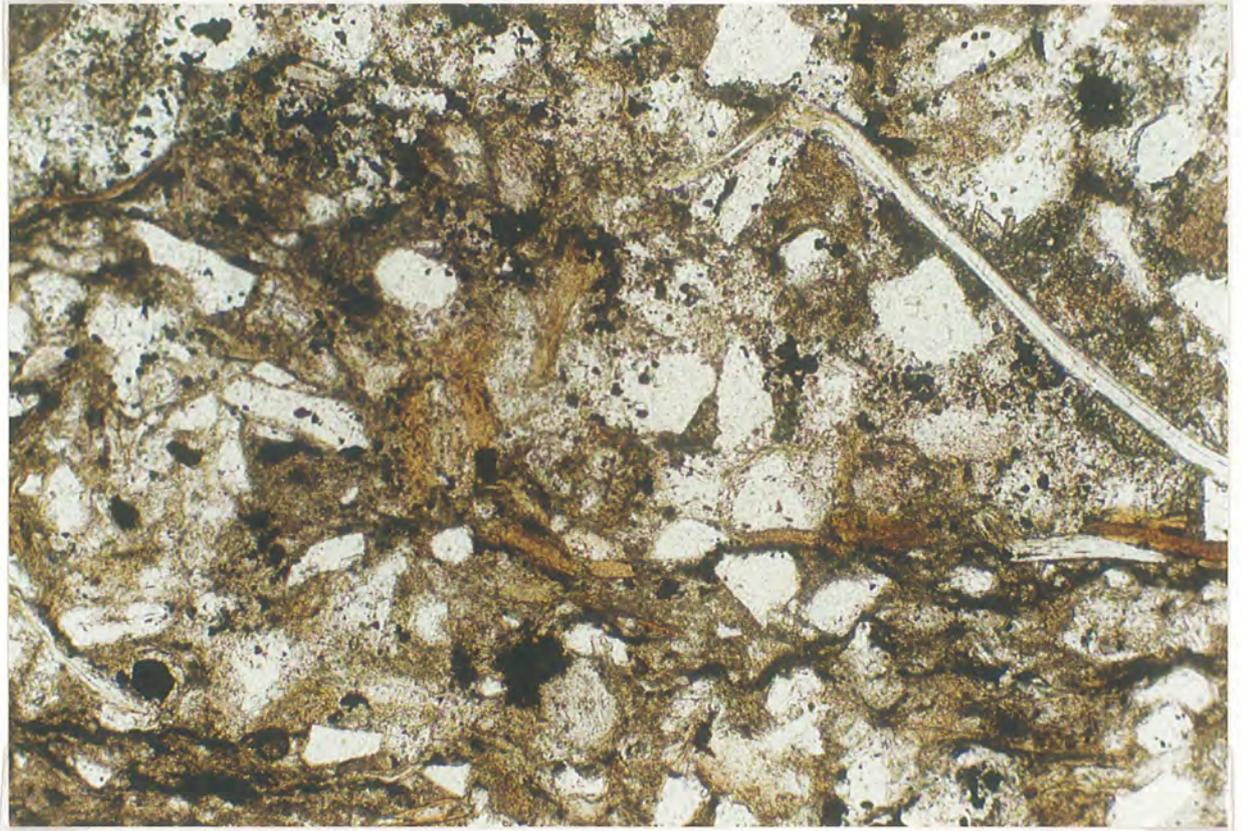
in turn.

These effects are most immediately obvious in the pelites of the Middle and Upper Pabba Beds. Outside the M1 aureole, these have a crude fissility or schistosity that is parallel, or very nearly so, to the well developed sedimentary lamination also present in these rocks. This schistosity is produced by a weak alignment of prominent fresh biotite and muscovite flakes up to 2mm long, and by closely spaced dark organic - rich stylolitic seams. The biotite in particular must be of metamorphic origin because it is not stable under diagenetic conditions ( see Chapter 5 ). A notable feature of many of the larger mica flakes is that they are bent around adjacent detrital grains ( Plate 3.53A ) or at the margins of stylolitic seams ( Plate 3.53B ). The strength of the preferred mica orientation varies greatly but is greatest in the stylolitic seams, suggesting that these have been produced by compaction and possibly by dissolution of more soluble components of the rock. The orientation of both the micas and the seams parallel to bedding indicates compression perpendicular to bedding. This predates the open concentric folds on the shore at Kilchoan which, as noted above ( section 3.2.5.4 ), fold the schistosity. This folding appears to be due to near - horizontal radial compression which can be attributed to the emplacement of the Hypersthene Gabbro and the associated doming ( see Chapter 7 ). One possible interpretation of the preceding syn - M0 vertical compression implied by the mica foliation and the stylolites is that it represents compaction of the original Jurassic mudstones, during low grade metamorphism, beneath the thick sequence of basalts and volcanoclastic rocks which formed early on in the history of the Ardnamurchan complex ( see Chapter 2 ). Alternatively, the schistosity, although formed by metamorphic biotites and muscovites, may be mimetic after aligned clay minerals in the original sediment, as appears to be the case in the illites described by Maxwell & Haver ( 1967 ). In this case the compaction may in fact be much older. The early age of formation of the schistosity is confirmed by the presence of aligned pseudomorphs after micas in rocks closer to the contact of the pluton which have been affected by high - grade M1 metamorphism ( Plate 3.54 ). M0 as recorded in these rocks certainly predates M1 and probably represents a more or less distinct earlier event. Unlike M1, it does not show an increase in grade towards the contact in these pelites.

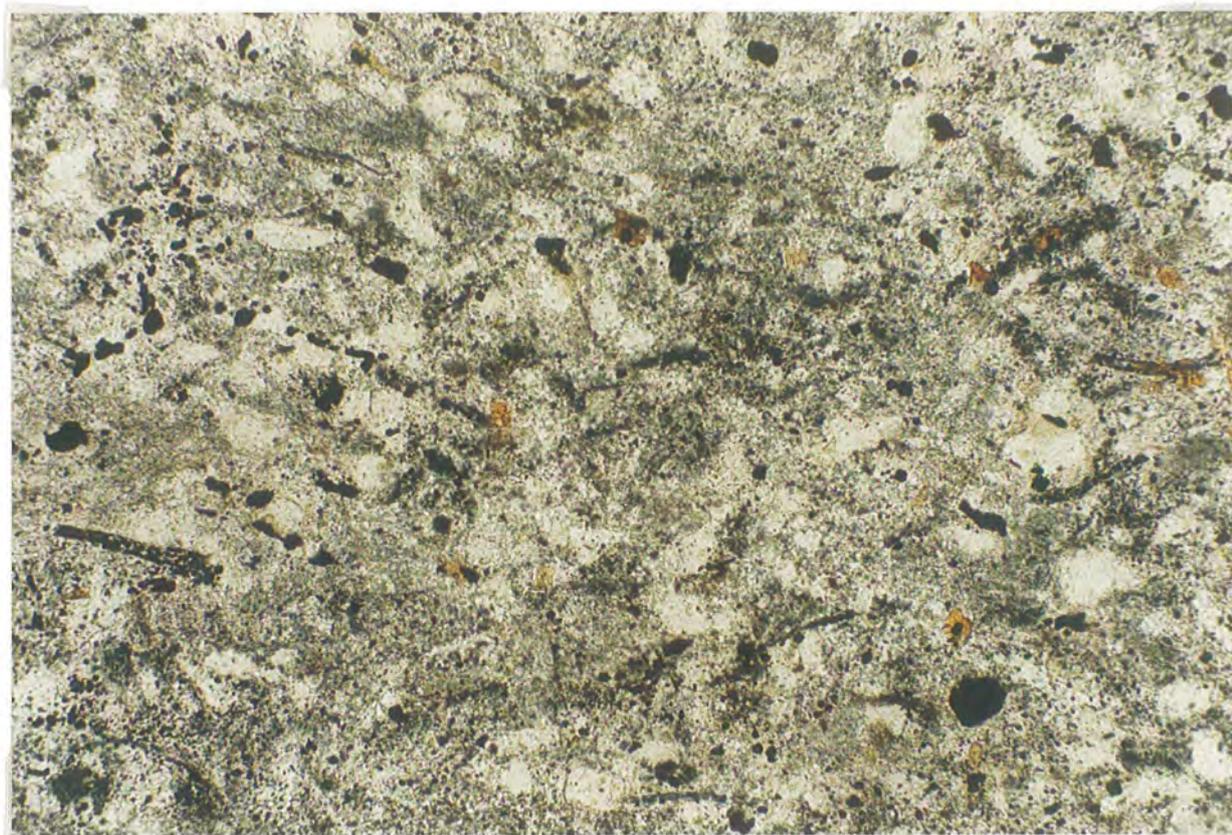
The Upper Pabba pelites in particular contain a number of originally calcareous inclusions, which include some nodular structures but are mainly shell debris and intact fossils such as *Gryphaea* and belemnite guards. These have been dissolved and replaced, even outside the M1 aureole, by coarsely

**Plate 3.53A.** Deformed M0 mica flakes defining weak bedding - parallel foliation in Middle Pabba Beds pelite from outside the M1 aureole. Sample 231. Plane polarised light, field of view 1.5mm.

**Plate 3.53B.** Stylolitic seam with relatively well - developed M0 mica foliation within the seam. Middle Pabba beds pelite ( sample 232/2 ) from outside the M1 aureole. Plane polarised light, field of view 3.5mm.



**Plate 3.54.** Opaque + K feldspar pseudomorphs after M0 micas. Sample 44E, a Middle Pabba Beds pelite from within the inner zones of the M1 aureole where aluminosilicates are absent ( see section 5.2.1 ). Plane polarised light, field of view 1.5mm.



crystalline calc - silicate assemblages ( epidote and, rarely, diopside or prehnite ) or by epidote + albite or adularia. The presence of these phases implies early carbonate loss from these initially carbonate - poor lithologies.

The originally more carbonate - rich rocks of the Middle Jurassic sequence on Maol Bhuidhe also show evidence of early decarbonation. Carbonate is in general absent except in a few calcite - rich, silica - poor rocks ( in which calcite may have been preserved simply as a result of insufficient silica with which to react to form calc - silicates stable under M0 conditions ) and in much later calcite ± dolomite + wairakite retrogressive assemblages in some calc - silicate rocks. Originally calcareous sandstones contain the assemblage quartz + clinozoisite + alkali feldspar ± actinolite ± sphene ± prehnite: diopside is present in some concretionary nodules in these rocks. The age of this pervasive decarbonation is indicated by early, pre - cone sheet veins in early alkali dolerite dykes and sills, and in the basalts at Sron Bheag, which contain various carbonate - poor assemblages. These include zeolites, prehnite, quartz + clinozoisite ± garnet + actinolite + sphene, clinozoisite or epidote + actinolite + albite, and only very rare prehnite + calcite veins. The host basic rocks, and an early felsite sheet at 45576263, also contain carbonate - poor alteration assemblages ( chlorite ± hornblende ± sphene + albite ± zeolites ). The presence of sphene is particularly significant because it is only stable in the presence of  $CO_2$  - poor fluids ( see Chapter 5 ). This implies that most of the Middle Jurassic rocks on Maol Bhuidhe, through which the fluids would have passed, had already undergone carbonate loss by the latter part of M0 at the latest, or were in the process of rapidly doing so. The only intrusions in the Maol Bhuidhe area which do show carbonate - rich alteration are the porphyritic felsites in the vent at Cuingleum and two composite felsite - basalt sheet intrusions further east, between 470627 and 468626. This indicates that these intrusions predate the decarbonation and were sealed to fluid flow during the period in which it took place. This would be consistent with these intrusions belonging to the early plateau basalt suite of rocks, as was suggested in Chapter 2.

There is actually little evidence to suggest that the decarbonation, unlike the schistosity in the pelites, predates M1: it is entirely possible that M0 in the calcareous rocks overlaps with the high - grade metamorphism further inland. All that can be said with certainty is that the decarbonation and associated veining must predate the cone sheets in the area, which consistently cross - cut the veins and do not show a similar type of alteration to that in the earlier metabasic rocks. These cone

sheets belong to Groups 1 and/or 2, and the decarbonation must therefore predate M2.

A final distinctive group of pre - M1 assemblages are found in amygdales in the early alkali basalts and picrites. In the M1 aureole these contain M1 assemblages such as aegirine - augite + olivine + feldspar, or aegirine - augite alone. These are similar in their bulk composition to the assemblages in amygdales outside the M1 aureole, such as zeolite  $\pm$  chlorite  $\pm$  saponite  $\pm$  carbonate, implying early M0 formation of these assemblages.

High temperature metamorphism on the southern margin of the Hypersthene Gabbro, as on its northern margin, is confined to a definite contact aureole within which grades of metamorphism rise sharply towards the contact. This is apparent from the presence of a number of mineral zones, particularly in the FeAl - rich pelites of the Middle Pabba Beds. The increase in grade towards the contact implies that, as on the northern margin of the pluton, the metamorphism was produced as a result of heating by the Marginal Border Group magma chamber ( see Chapter 5 ). The episodes of high grade metamorphism on both northern and southern sides of the pluton therefore seem to be the same M1 metamorphism, although this does not necessarily imply that M1 was entirely synchronous around the pluton. The limit of of recognisable M1 metamorphism lies as much as 700m from the contact on the southern side of the pluton, in contrast to a limit of 300m on the northern margin at Glendrian Bay. This may be the result of a longer period of conductive heating, between initial emplacement of the MBG magma chamber and the end - M1 quenching event, on the southern margin than on the northern margin, although other interpretations are possible ( see section 5.2.4 ). It should be noted that the outer contact of the MBG is very steeply inclined on both northern and southern margin so the difference in aureole width is not simply an artefact of exposure.

The outer limit of M1 metamorphism on the southern margin is marked by the near - coincident disappearance of M0 biotite and muscovite in the pelites, and their replacement by very fine - grained granular K feldspar - cordierite - magnetite - aluminosilicate aggregates ( see Plate 3.54, above ). A second reaction, observed around 300m from the contact, involves the appearance of granular orthopyroxene and a gradual reduction in amounts of cordierite and opaques. Within about 200 metres of the contact, the pelites show evidence for partial melt formation in the shape of interstitial quartz - feldspar intergrowths and small diffuse quartzofeldspathic segregation veinlets,

which contain accessory poikilitic orthopyroxene ( Plate 3.55 ). Contortion of relict compositional lamination is often visible in thin sections of these rocks, indicating small - scale rheomorphism ( Plate 3.56 ).

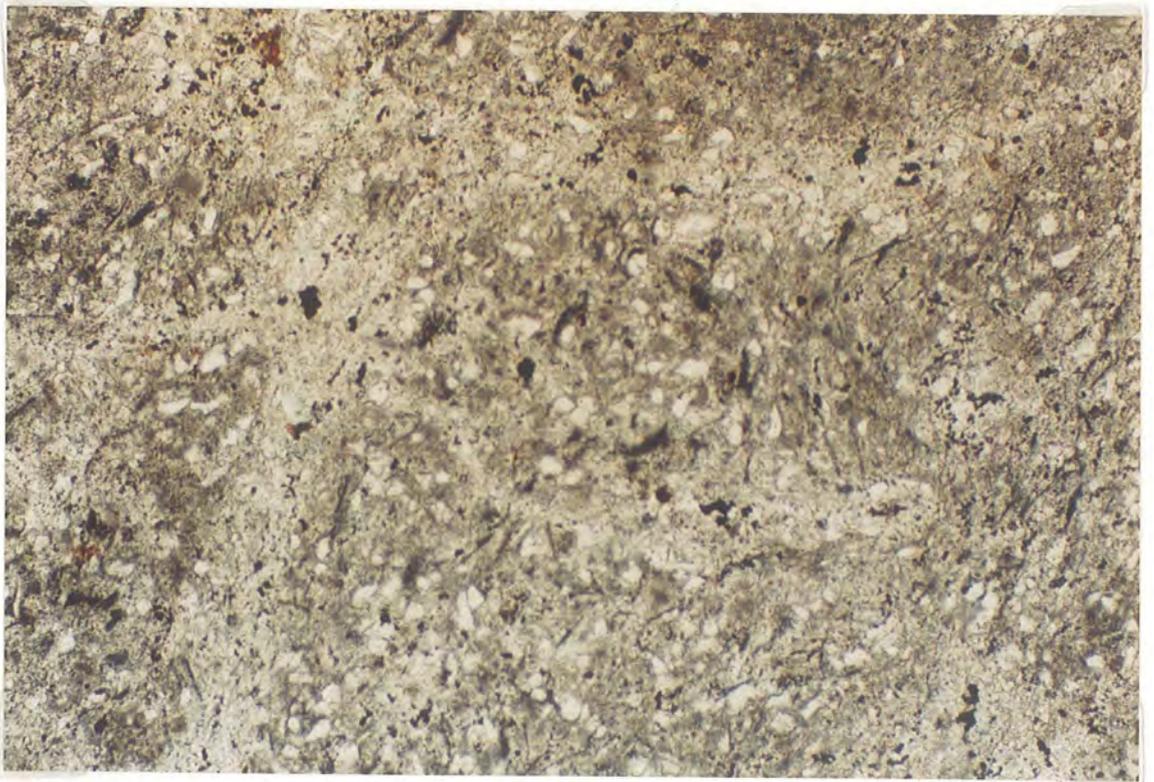
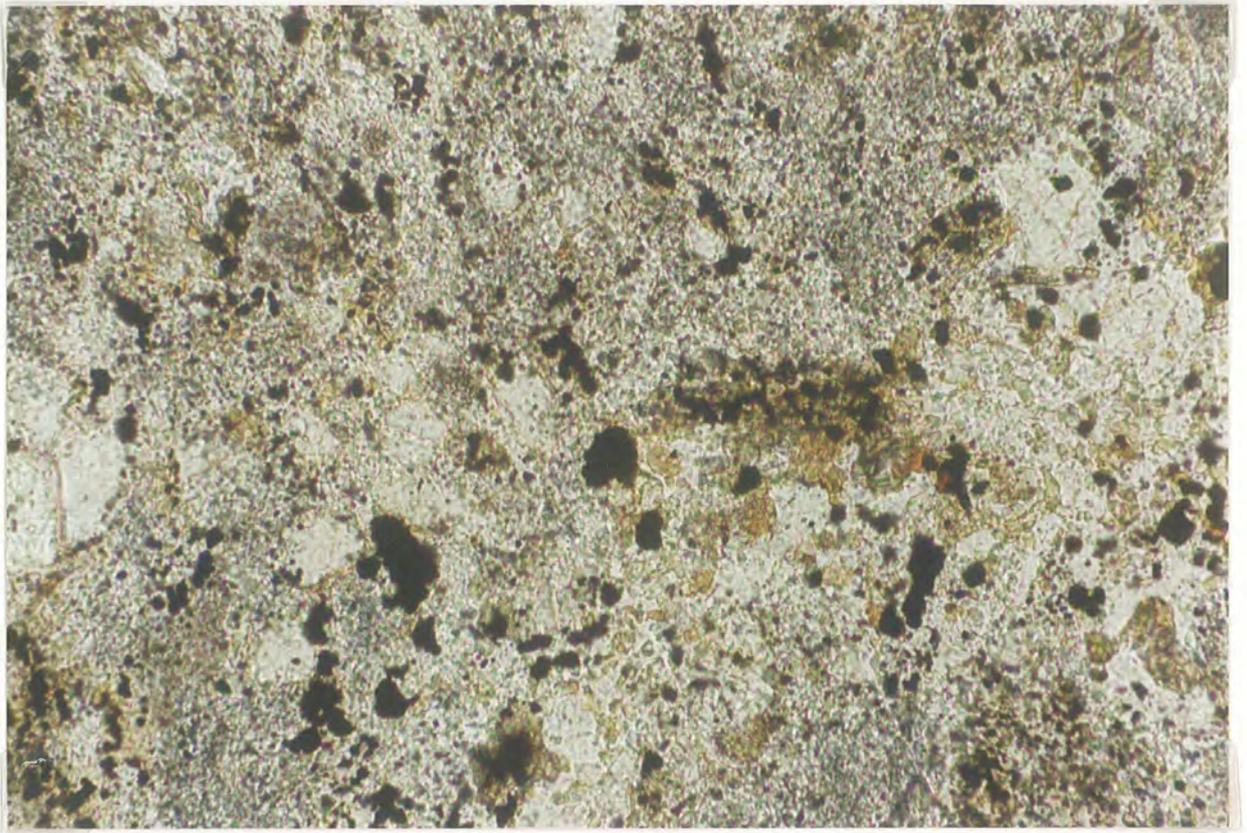
Related high - grade assemblages occur in the ferruginous siltstones of the Lower Pabba Beds. These are not accessible in exposures more than 300m from the contact so the lower grades of M1 metamorphism could not be studied in them. However, the samples collected do contain identical pseudomorphs after micas to those found on the pelites, indicating similar mica breakdown reactions. Cordierite is present but is rare: the dominant ferromagnesian silicate is orthopyroxene, and magnetite is also abundant. Most samples of the Lower Pabba Beds contain an interstitial granophyric to granular intergrowth of alkali feldspar and poikiloblastic quartz, indicative of recrystallisation of a quenched felsic melt ( Lofgren 1971 ). The quartz in this material commonly pseudomorphs clusters of prismatic grains which may have been quench tridymite ( Plate 3.57 ). In a few outcrops only, evidence of mesoscopic melt mobility was observed, in the form of unchilled granular microgranitic veins which root out into the more quartzose beds and are only visible where they cut the more pelitic laminae ( Plate 3.58 ).

In contrast to the relatively well preserved M1 pelitic assemblages, the calcsilicate rocks of probable Middle Jurassic age on Druim na Gearr Leacainn, and even many of the calc - pelites of the Upper Pabba Beds, are extensively retrogressed. Diopside  $\pm$  hornblende  $\pm$  garnet  $\pm$  idocrase assemblages are typical: as will be shown in Chapter 5 these are too low - grade to have formed during M1 and are probably M2 in age, although direct evidence from age relationships is lacking.

Pre - M1 basic igneous rocks only occur in the inner part of the aureole, and mainly only in the downfaulted blocks, although originally doleritic sills are exposed south of the concentric fault system in the floor of the Dubh Chreag gorge around 45436318. Although originally transitional to alkaline, the basalts in particular were affected by pre - M1 alteration ( see Chapter 2 ). As a result of this orthopyroxene is a common phase in the fine grained granular to poikiloblastic groundmass of these rocks, along with opaques, augite and plagioclase. Granular opaque oxide grains are particularly abundant, indicating oxidation of these rocks during alteration or possibly during M1. Olivines and titaniferous augites of the primary igneous assemblage in these rocks are wholly or partly replaced by granular pyroxene - opaque aggregates, whilst primary feldspars are corroded and contain numerous

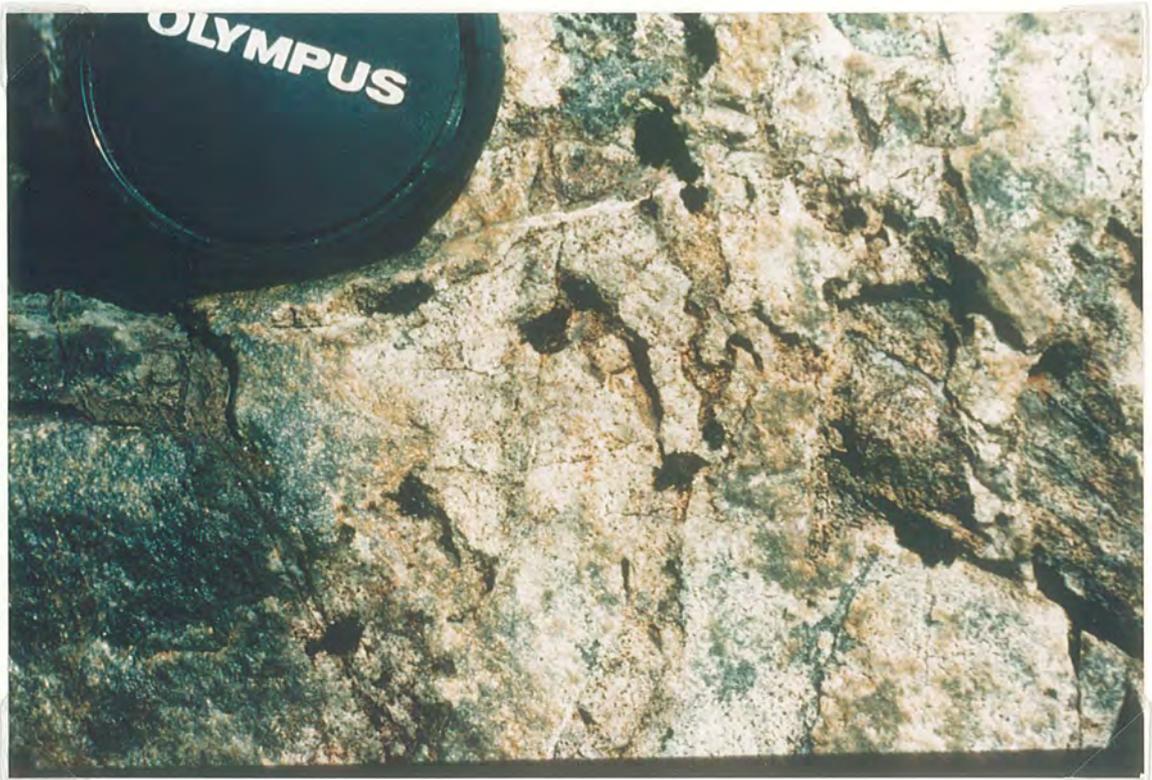
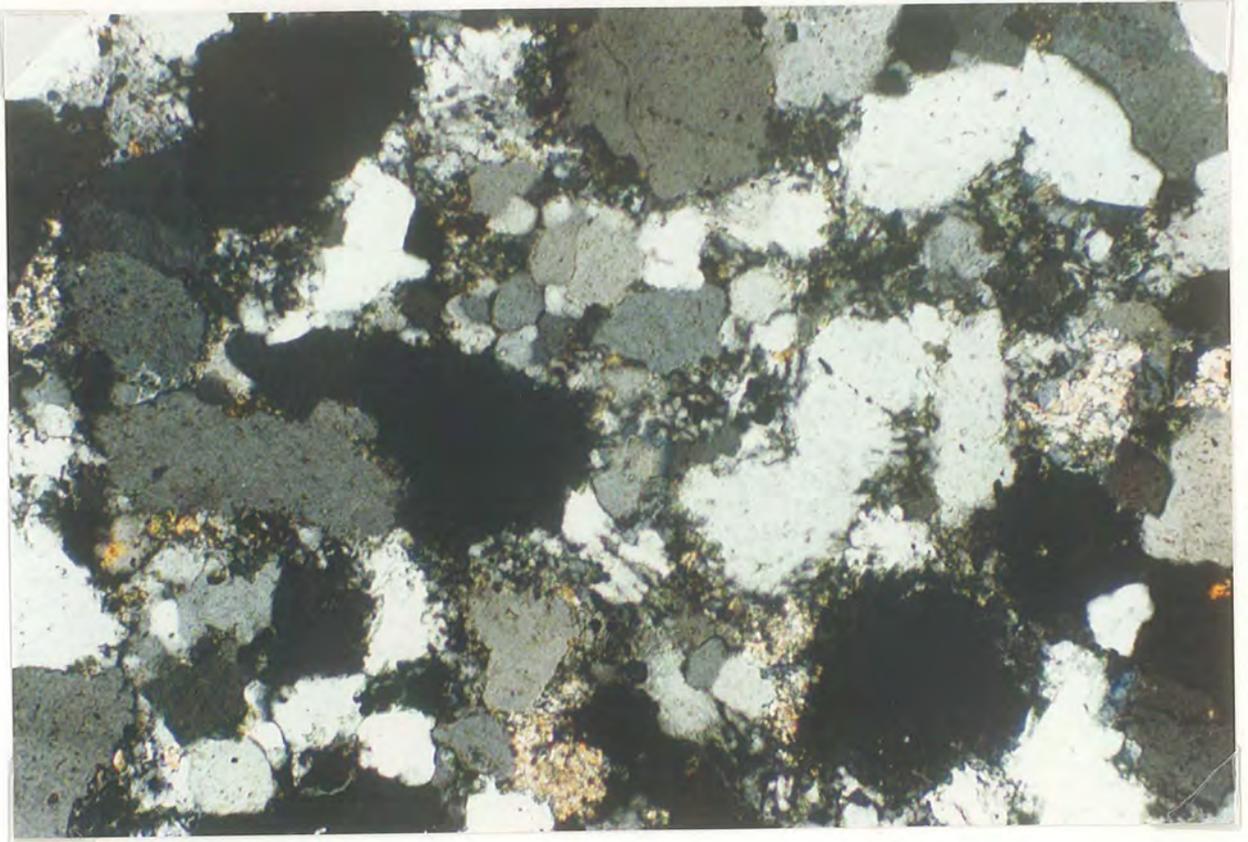
**Plate 3.55.** Diffuse M1 quartzofeldspathic melt segregation vein within Middle Pabba Beds pelitic hornfels. Vein contains accessory poikilitic orthopyroxene. Sample 87/2. Plane polarised light, field of view 0.75mm.

**Plate 3.56.** Diffuse segregation veins in Middle Pabba Beds pelitic hornfels ( sample 44E ). Relict lamination at the right - hand side of the field of view is contorted near the segregation veins. Plane polarised light, field of view 3mm



**Plate 3.57.** Partially fused quartzose semi - pelite, Lower Pabba Beds ( sample 120F ). Poikiloblastic quartz in the centre of the field of view replaces stubby prisms ( possibly quench tridymite ? ) in interstitial quartz - feldspar intergrowth. Crossed polars, field of view 1.5mm.

**Plate 3.58.** Unchilled quartzose microgranitoid vein cutting pelitic laminae roots out into the quartzose beds on either side. Partially melted Middle Pabba Beds hornfelses. Field location 120B2, grid reference 45456318.



granular inclusion trails of pyroxenes and opaques ( Plate 3.59 ): these may correspond to fractures, filled with chloritic material, which occur in dolerite sills affected by M0 which occur on the coast south - west of Maol Bhuidhe.

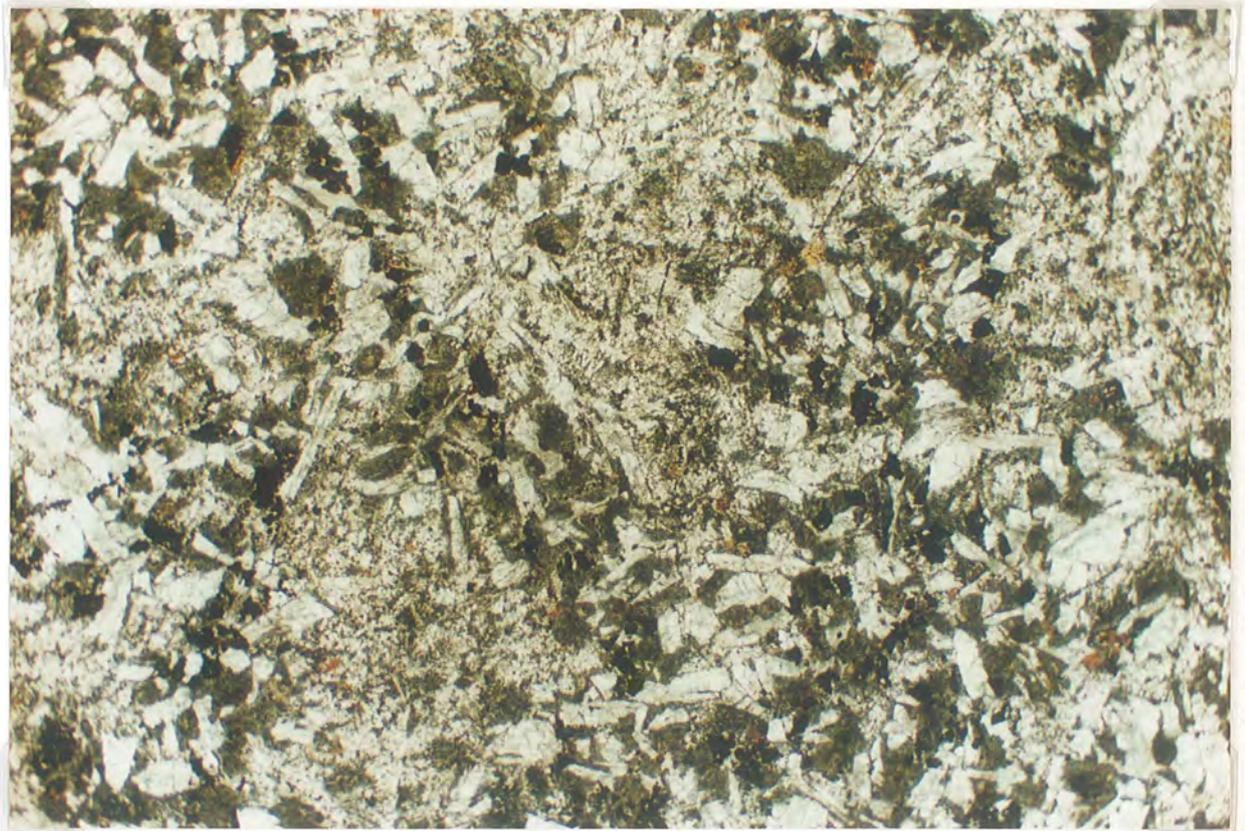
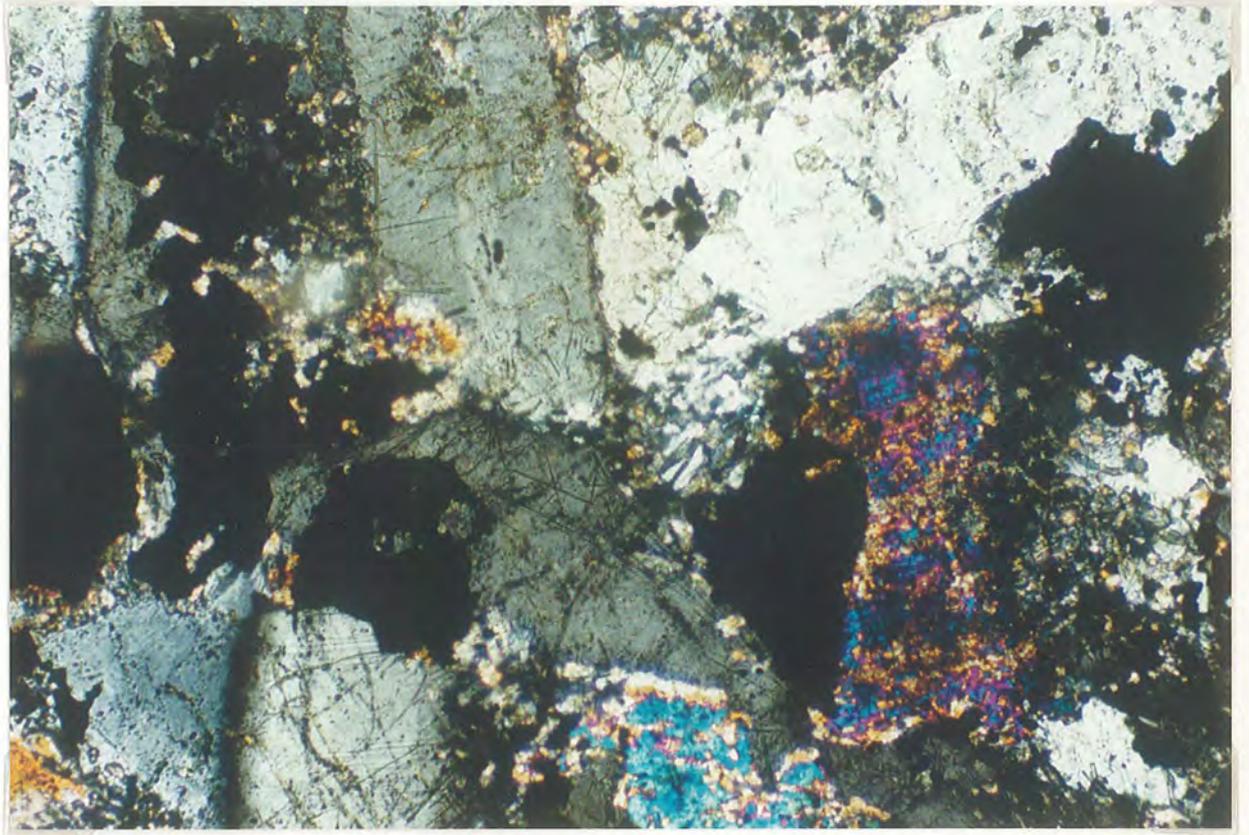
The degree of recrystallisation observed in these early metabasic rocks shows an overall increase with composition from the picrites, which are almost pristine except for areas of previous alteration which are completely recrystallised, to fractionated interstitial material in the dolerite sills. This interstitial material is completely recrystallised, to a granular aggregate of plagioclase, alkali feldspar, opaques, pyroxene and apatite. There is, however, little evidence for actual melting in these rocks, except in the rare amygdales which have not been replaced during M2. These contain coarse poikilitic aegirine - augite, indicative of crystallisation from a melt. The strong contrast between the advanced state of recrystallisation in the more evolved basalts and alkali dolerites, and the relatively well - preserved igneous textures in the cone sheets noted by H.H.Thomas ( in Richey et al. 1930 ) was also found to be a general feature of the rocks studied in the course of this work, although unfortunately these did not include cone sheets from the An Acairseid - Kilchoan sector which were known to be cross - cut by the outer contact of the MBG. The few examples of such sheets found, for example at field loc. 183D, do have a granular appearance in hand specimen and would appear to have been affected by M1, but this would need to be confirmed by detailed petrography.

Although the metabasic rocks in particular contain diffuse recrystallised granoblastic - textured veins, there are no syn - M1 veins in these rocks with textures analogous to those of the high - temperature hydrothermal veins observed in the Inner Series of the Hypersthene gabbro ( section 3.4.3 and Chapter 6 ). It appears that there was no network of fractures open during M1 which could have acted as pathways for vigorous convective hydrothermal circulation. This implies that such flow was unlikely to have developed during M1 in the high - grade aureole, because this requires a high permeability, which in such strongly indurated rocks would most probably have to be fracture - based ( Norton 1984 ). As is the case on the northern side of the pluton, M1 on the southern margin seems to have been a largely thermal metamorphic event.

Another feature common to M1 as developed on both exposed margins of the pluton is that it ended with the quenching of partial melts and high - grade mineral assemblages. Part of the reason for the poor preservation of quench textures, particularly quench glass textures, on the southern, relative

**Plate 3.59A.** Relict igneous plagioclase laths in pre - M1 alkali dolerite ( sample 120C ) with granular inclusions and inclusion trails of pyroxene and opaques. These may be the hornfelsed remnants of chlorite - filled fractures similar to those found in alkali dolerites outside the M1 aureole. Crossed polars, field of view 1.5mm.

**Plate 3.59B.** Large - field photomicrograph of 120C showing partly preserved doleritic texture and patches of granular felsic material a few millimetres across which may be the hornfelsed ( and possibly remelted ? ) remains of residual melt segregations in the original rock. Plane polarised light, field of view 14mm.



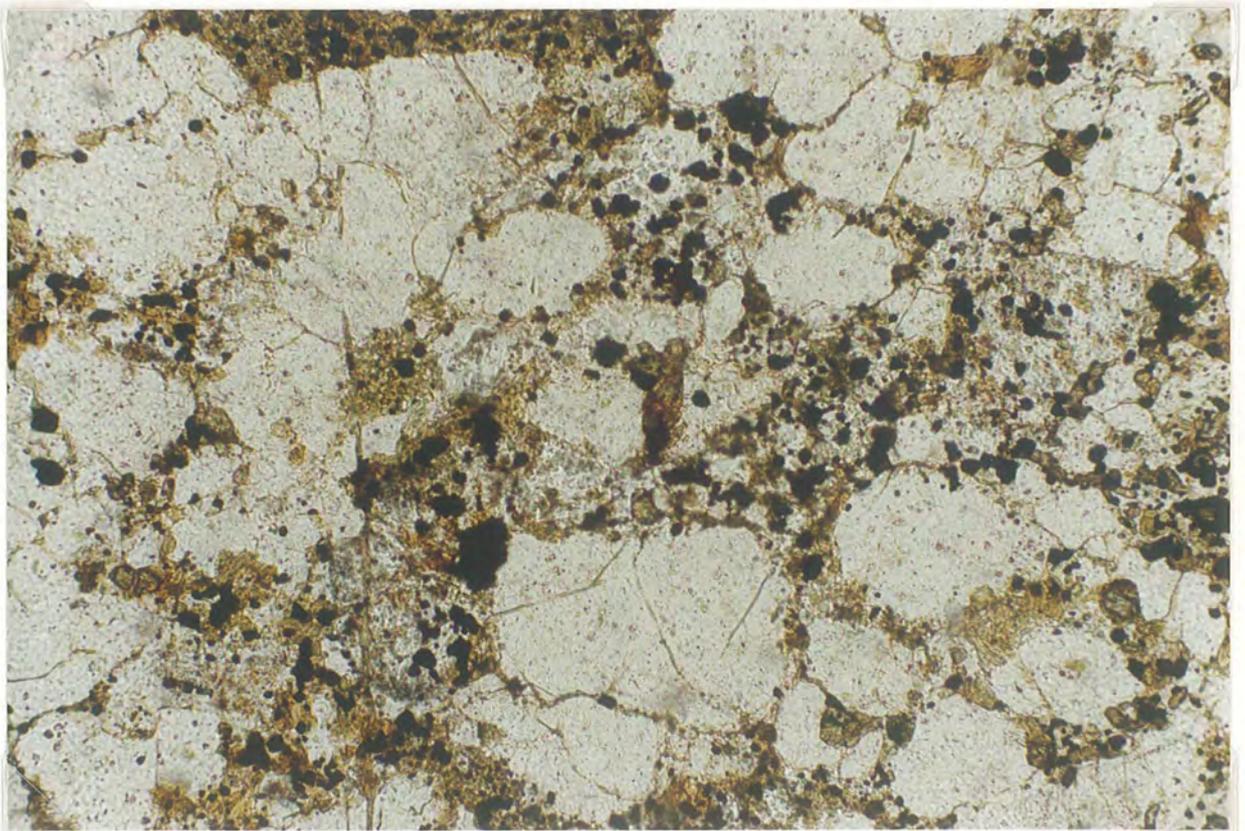
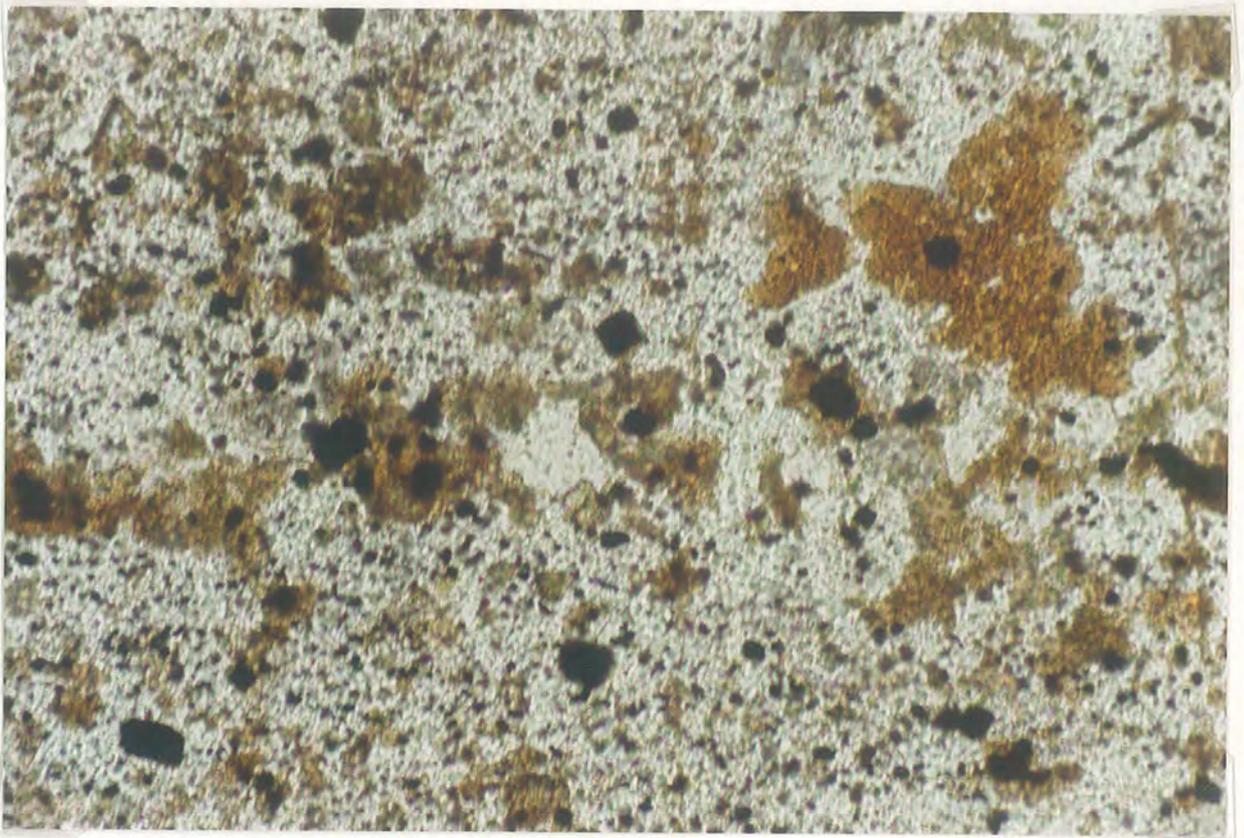
to the northern margin, and especially in comparison with the preservation of these textures in the Glendrian Bay area ( section 3.2.2 ) may be the greater intensity of M2 on the southern margin. This is expressed both in more complete overprinting of M1 assemblages and in the generally higher metamorphic grades of the M2 assemblages close to the contact.

Like the M1 thermal aureole, M2 metamorphism in this area shows a marked zonation which is best developed in the FeAl - rich pelites of the Middle Pabba Beds. It is, however, complicated by the retrogressive character of M2, which has resulted in the earlier, higher - grade M2 assemblages being themselves partially overprinted by lower - grade M2 assemblages. Thus an initial partial replacement of the opx - bearing M1 assemblage, close to the contact, by biotite + cordierite + K-feldspar + plagioclase + quartz + magnetite ( Plate 3.60 ), is itself altered, with the replacement of biotite by chlorite and cordierite by pinitite. At distances of more than 200m from the contact, chlorite - bearing M2 assemblages are ubiquitous in the pelites and M2 appears to have been low - grade throughout. Similarly, early M2 hornblende and biotite in the Lower Pabba Beds ferruginous siltstones are replaced by chlorite and epidote ( Plate 3.61; a notable feature of this photomicrograph is the prominent amphibole filled tensile fracture visible in it ). Later, lower grade M2 assemblages are particularly well developed in the more calcareous Upper Pabba Beds and in the probably Middle Jurassic rocks on Druim na Gearr Leacainn, with earlier grossular, idocrase and clinopyroxene being replaced by actinolite and epidote bearing assemblages. The assemblage epidote + actinolite + albite or adularia is particularly common in these rocks, replacing fossils or filling fractures. Further from the contact, on Maol Bhuidhe and on the coast at Kilchoan , even lower grade assemblages are present, in the form of calcite + dolomite + wairakite replacement of earlier calc - silicate assemblages, and calcite - rich veins in faults. In contrast, the pre - M1 metabasaltic and metapicritic rocks of the plateau basalt suite ( section 2.4 ) commonly preserve relatively pristine M1 assemblages, although early M2 red amphibole ( probably a pargasitic hornblende or oxyhornblende ) commonly rims and corrodes magmatic or M1 olivine and granular M1 orthopyroxene in these rocks.

The progressive evolution of M2 assemblages developed in these older rocks, together with a general decrease in the intensity of alteration in the rocks concerned, allows a subdivision of the cone sheets in this sector of the M2 aureole which is consistent with the age relationships of the groups of cone sheets defined in section 3.2.5.4:

**Plate 3.60.** Middle Pabba Beds pelite showing replacement of M1 orthopyroxene and Ti - magnetite by M2 poikiloblastic biotite. Sample 177/1. Plane polarised light, field of view 0.75mm.

**Plate 3.61.** Lower Pabba beds semi - pelite showing partial replacement of M1 orthopyroxene and Ti - magnetite by M2 biotite and green amphibole. Note partially annealed, partly amphibole - filled tensile fracture at left - hand side of plate. Sample 120B2. Plane polarised light, field of view 0.75mm.



**Plate 3.62.** High grade M2 metamorphic assemblage in early M2 cone sheet dolerite. Note oxidation of augite, with production of opaque exsolution lamellae and granular opaque inclusions, and local replacement of titaniferous magnetite by biotite. Sample 55/2. Plane polarised light, field of view 1.5mm.



1) Group 2 cone sheets and associated minor intrusions show replacement of any magmatic orthopyroxene which may have been present by green amphibole and/or chlorite; recrystallisation of primary augite ( characterised by the presence of exsolution lamellae ) to granular augite + opaque aggregates, or to subophitic grains lacking exsolution lamellae but containing abundant opaque inclusions; patchy partial replacement of augite by green hornblende and/or chlorite in all samples more than about 200m from the contact, and in most others; complete replacement of olivine, when originally present, by talc + opaques; recrystallisation of any felsic interstitial material to rather granular quartz + alkali feldspar + plagioclase aggregates; and replacement of primary titaniferous opaques in rocks close to the contact by red Ti - rich biotite ( Plate 3.62 ). One section contains small interstitial patches of granular orthopyroxene which may be M2 in age. These rocks show spatial and temporal variation in the products of M2 metamorphism like that seen in the pre - M2 rocks, implying that they were emplaced during the earlier part of M2.

2). Hydrothermal alteration in the Group 3 cone sheets is of relatively low grade at all distances from the contact with the Marginal Border Group: chlorite, green hornblende and epidote are the characteristic metamorphic phases. This implies that these sheets were emplaced during the latter part of M2.

3). The various intrusions of Group 4 show even less alteration than Group 3, with chlorite and sericite being the typical products of alteration in these rocks where they occur outside the pluton.

Although the overall pattern of variation in M2 is simple, with the higher grade assemblages forming earlier and closer to the MBG contact, in detail the pattern is more complex. Estimation of actual conditions of metamorphism is in part dependent on mineral composition data and is dealt with in Chapter 5, but variations in the intensity of alteration are strongly developed and are easily mapped in the field. On the hand - specimen and thin section scale, M2 assemblages are most strongly developed in alteration haloes around veins, and a similar pattern is also developed on much larger scales. The later, lower - grade alteration in particular is most intense along and around radial faults. Examples of faults associated with intense late M2 alteration include the Dubh Chreag gorge fault, the fault on the western side of Tom na Moine, and the complex fault system around An Acairseid ( Map 4 ). All of these cut the Inner Series and some cut later intrusions as well, which raises the possibility that at least part of the later alteration of the rocks around them may have

been caused by later intrusions and that the later parts of M2 therefore postdate the Hypersthene Gabbro. Alteration is also concentrated along the concentric fault systems, but to a much smaller degree. This may be because of compressive stresses across these faults, produced by the weight of the overlying rocks, which would tend to close up planar fractures within them.

To summarise, the history of metamorphism along the southern margin of the Hypersthene Gabbro shows many similarities to that on the northern margin: early high to very high grade thermal metamorphism ( M1 ) following initial emplacement of the intrusion; rapid quenching of the contact aureole and the contact zone as well; subsequent hydrous, lower - grade ( M2 ) metamorphism associated with a joint - like fracture and vein network.

#### **3.2.5.6. The Problems of Spatially Limited and Temporally Extended Periods of Activity in the determination of Age Relationships on the southern margin of the Hypersthene Gabbro.**

Extensive use has been made in this section of two indirect methods of determining relative ages and correlating events in widely separated outcrops. These are the use of cross - cutting relationships involving petrographically distinctive suites of cone sheets and other minor intrusions, particularly the Group 3 cone sheets, and of overprinting by distinctive metamorphic assemblages. Such methods have been widely applied in unravelling the histories of regional metamorphic terranes, but their use in a complex contact metamorphic setting involves certain additional problems which are considered here.

The use of the suites of cone sheets and other minor intrusions as time markers assumes that their source magma chamber(s) changed composition with time alone and were compositionally uniform at any one time. A particular problem arises in the case of zoned source magma chambers, because these will tend to emplace cone sheets of different composition, originating from different levels of the magma chamber, in different areas of the country rocks, for reasons related to the mechanism of emplacement of the cone sheets which are discussed in Chapter 7. One way of testing the assumption of a homogenous source magma chamber is by geochemical comparison of minor intrusions with rocks of known age relationships within the parent intrusion(s), in this case the

Hypersthene Gabbro and/or other intrusions within Centre 2 of the Ardnamurchan complex. A second method is to test the age relationships deduced using the minor intrusion suites method for consistency with age relationships deduced by other means, such as the use of phases of deformation and metamorphism. Unfortunately, these suffer from their own particular problems when applied to the southern margin of the Hypersthene Gabbro.

The principal problem encountered when using the phases of metamorphism on the southern margin of the pluton as time constraints is that the metamorphism shows strong spatial variation, especially during M1, which is not developed at all more than 700m or so from the contact ( see section 3.2.5.5, above, and Chapter 5 ). Thus although M0 and M2 can be distinguished, and rocks and structures dated relative to them, at greater distances from the contact than the outer limit of M1, the age relationships of pre - M2 rocks and structures, as well as M0 metamorphic phenomena, relative to M1 can only be determined by correlation with similar structures and rocks within the M1 aureole: an example of this is the dating of the M0 foliation in the Pabba Beds. The spatial variation in M2 could cause similar problems although the clear age relationships between M2 and the Group 2 and Group 3 cone sheets close to the contact, and the use of these very distinctive groups of cone sheets as time markers elsewhere, provides a cross - check on the use of M2.

A further problem associated with the use of phases of metamorphism and suites of minor intrusions is that the metamorphism or the emplacement of the intrusions will in general take place over an extended period of time and consequently will not provide precise time constraints. One example is the age relationships of the doming around the Hypersthene Gabbro, which predates M1 but postdates much of M0, and the syn - M0 faulting around Maol Bhuidhe. It is not possible to tell from the age relationships of these to the metamorphism whether the faulting at Maol Bhuidhe predates the doming or not. Similarly, the inward - dipping concentric normal and reverse fault sets cannot have been active simultaneously but the time - separations of the activity cannot be resolved using the age relationships shown in Fig. 3.18. It should be noted that Fig. 3.18 has been constructed on the assumption that the M1 to M2 transition was synchronous outside the pluton, as on previous diagrams of this type, but the possibility that M1 and the later stages of M0 may have overlapped has been allowed for.

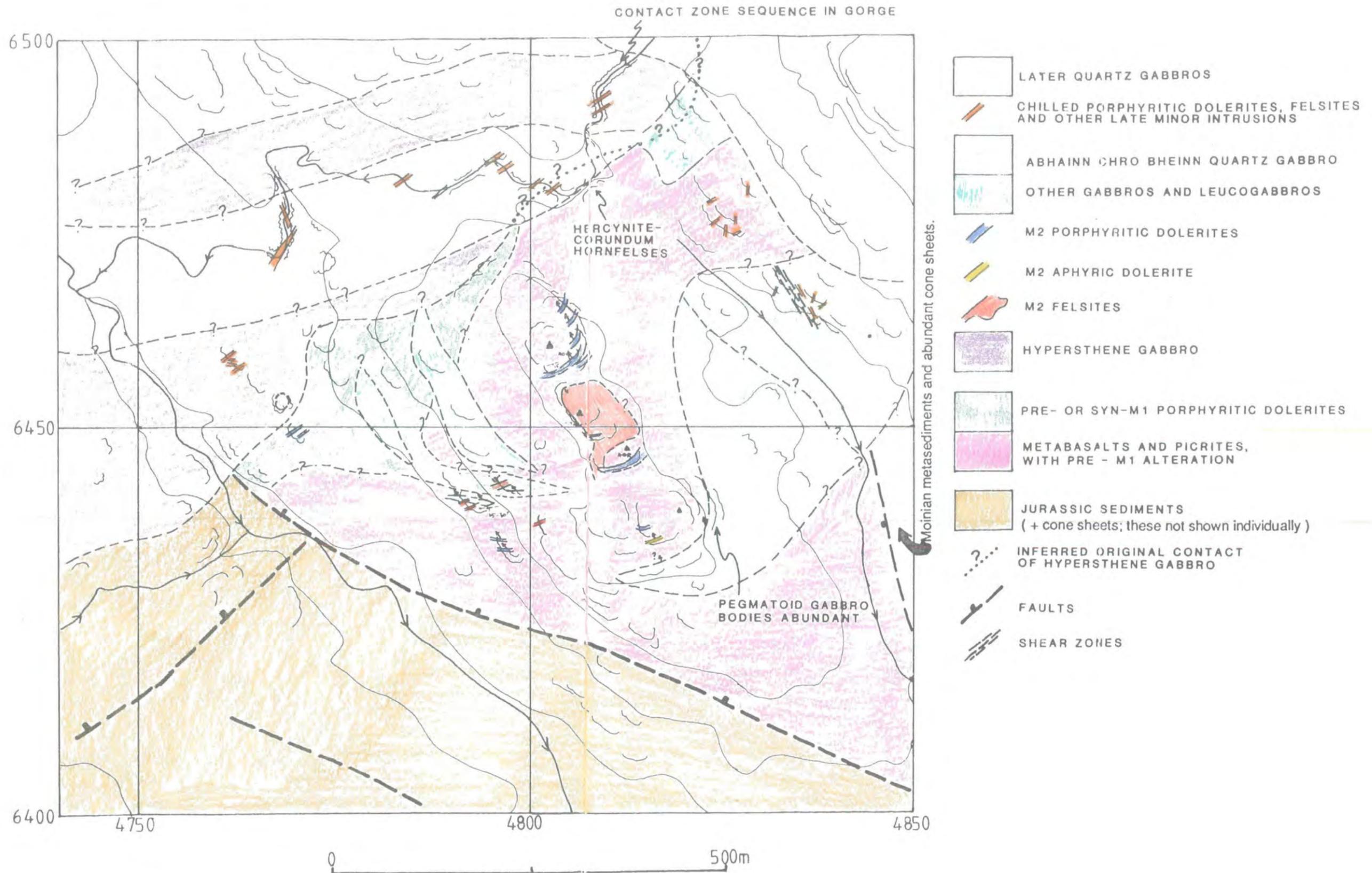
### **3.2.6. Glebe Hill and the Glas Eilean fault.**



The area considered in this section lies between two faults, the Kilchoan Bay fault to the west and the Glas Eilean fault to the east and south ( see Map 1 ), and consists mainly of a downfaulted block of Tertiary basalts and picrites intruded by cone sheets. On Glebe Hill itself these are intruded by gabbroic rocks which occur in a few outcrops on the western side of the hill around 47966442; in the valley of the Abhainn Chro Bheinn to the north; on the eastern and southern side of the hill around 48146456; in poor outcrops in the floor of the valley to the east of the hill, and on the eastern side of this valley. The area was therefore previously mapped with a gently dipping roof contact between the gabbros, believed to be part of the Hypersthene Gabbro, and older overlying basalts and cone sheets ( Richey et al. 1930, Wells 1954 ). However, the age relationships, petrography and geochemistry of many of the gabbros in this area suggest that they may not be part of the Hypersthene gabbro at all. The first two lines of evidence are dealt with here, and the geochemical evidence is considered in Chapter 6. A revised map of the area is included in this section as Fig. 3.19.

Rocks which do appear to be part of the Hypersthene Gabbro are, however, found to the north of Glebe Hill, especially in a small gorge cut by the Abhainn Chro Bheinn between 48096487 and 48146495, close to the contact with later quartz gabbro intrusions. The lower part of the gorge exposes fine - grained, homogenous quartz - poor gabbros. From 48126492 to about 48136493, heterogenous mottled rocks are exposed. These range from microgabbros through coarse mesocratic rocks to veins and irregular diffuse patches of granitic rock. Some of the granitic patches have fine - grained granophyric cores. At the very top of the gorge rounded and chilled pillows and sub - angular blocks of dolerite occur in a coarser dioritic host with pods and veins of granophyric - textured granitic rock. The dolerite contains plagioclase porphyrocrysts ( possibly xenocrysts ) and small, very coarse - grained xenoliths of anorthositic rock cut by granular annealed shear zones. Apart from the anorthosite xenoliths, these rocks, and the sequence in which they are exposed, proceeding up the gorge, are similar to that found in the contact zone of the Marginal Border Group at, for example, the eastern side of An Acairseid. Since no other intrusion in Ardnamurchan has a comparable suite of rocks at its outer margin, these rocks are believed to be part of the Hypersthene Gabbro: their age relationships, as described below, are consistent with this but do not prove it. If this identification is correct, it has the implication that the original outer contact of the Hypersthene Gabbro ( since cut out by later intrusions ) was only a few tens of metres at most from the top of the gorge, and probably faced east or north - east, suggesting that there may have been a very sharp

FIG. 3.19. REVISED MAP OF THE GLEBE HILL AREA SHOWING NUMEROUS SMALL POST - HYPERSTHENE GABBRO INTRUSIONS.



bend in the contact in this area.

The basic rocks in the gorge have a moderately granular, hornfelsed texture, with the assemblage plagioclase + augite  $\pm$  orthopyroxene + opaques. The low abundance or absence of orthopyroxene may have been caused by oxidation during metamorphism, since the augite in these rocks contain abundant opaque oxidation - exsolution lamellae and opaques are very abundant ( see Chapter 6 for further discussion of the reactions involved in such a process ). In contrast the felsic rocks are relatively unmetamorphosed and commonly contain elongate quench augite crystals and interstitial patches of coarse granophyre. A possible explanation of this is that the felsic rocks were partially molten during the phase of metamorphism recorded by the mafic rocks with which they are mingled, although in this case textural evidence of relict phases in the felsic rocks would be predicted: this is not present. The felsic rocks cannot, however, have been intruded into the mafic rocks during this phase of metamorphism because the presence of pillowed contacts between the two rock types implies that the two were originally emplaced together as magmas. All of these rocks show low - temperature hydrous alteration, with replacement of orthopyroxene by chlorite, replacement of augite by green hornblende, albitisation and epidote formation in plagioclase, and formation of drusy quartz, chlorite and calcite in cavities. Given the presence of similar assemblages in later rocks ( see below ) and the proximity of several later large intrusions, this hydrous metamorphism is probably a much later phenomenon and cannot be directly correlated with M2. If the rocks in the gorge are part of the Marginal Border Group, M2 in this area must be represented, at least in part, by the high grade oxidizing metamorphism recorded in them.

Comparable hornfelsed fine - grained gabbros, which are also magnetite - rich and orthopyroxene - poor, occur further west, in the bed of the Abhainn Chro Bheinn at 47686482 and on the hillsides to the north and south of this stream ( Fig. 3.19 ). An outcrop at 47936469 exposes an unchilled, intricately veined and locally diffuse contact between hornfelsed microgabbro and granoblastic basic hornfelses of the porphyritic group ( defined below ). In thin section, both rocks show granoblastic texture but this is less completely developed in the microgabbro, which retains relics of a primary ophitic texture.

The petrography of these rocks is consistent with their being metamorphosed equivalents of the Marginal Border Group of the Hypersthene Gabbro although their age relationships to the latter are

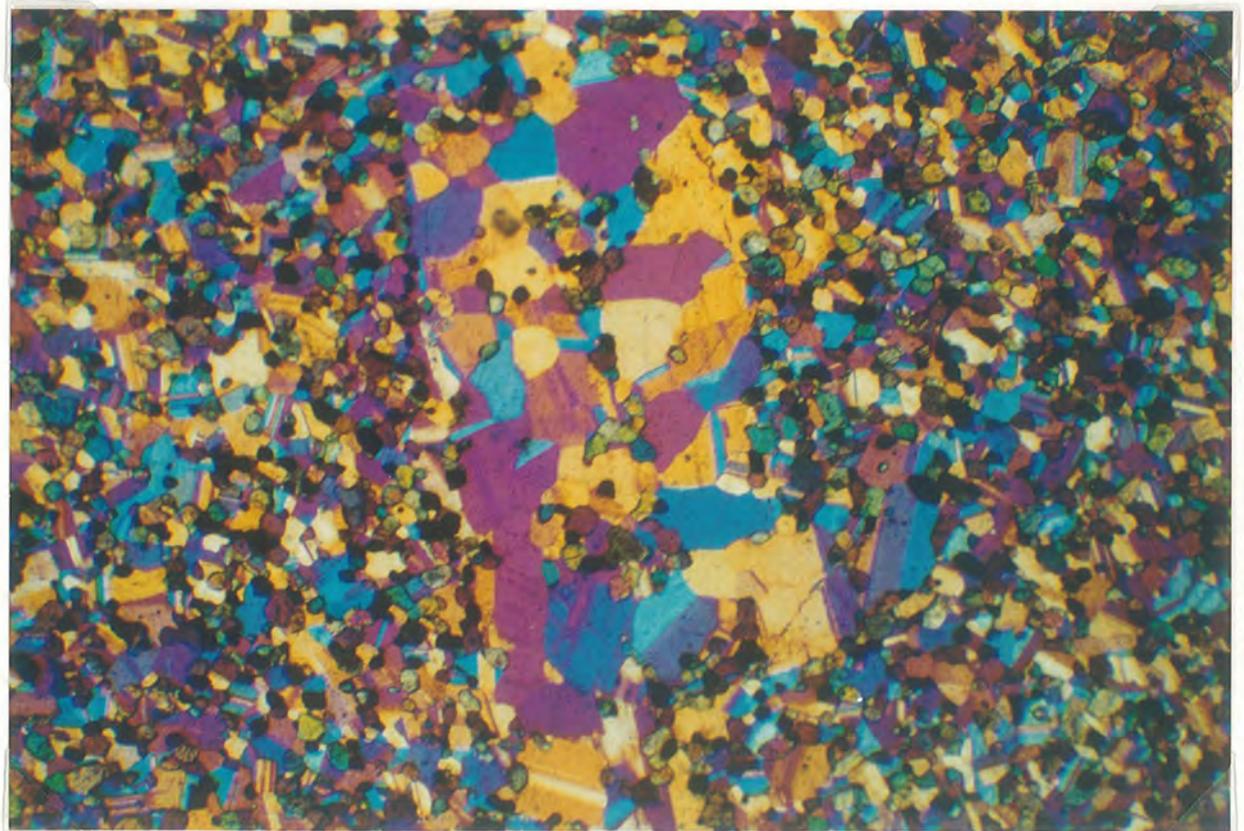
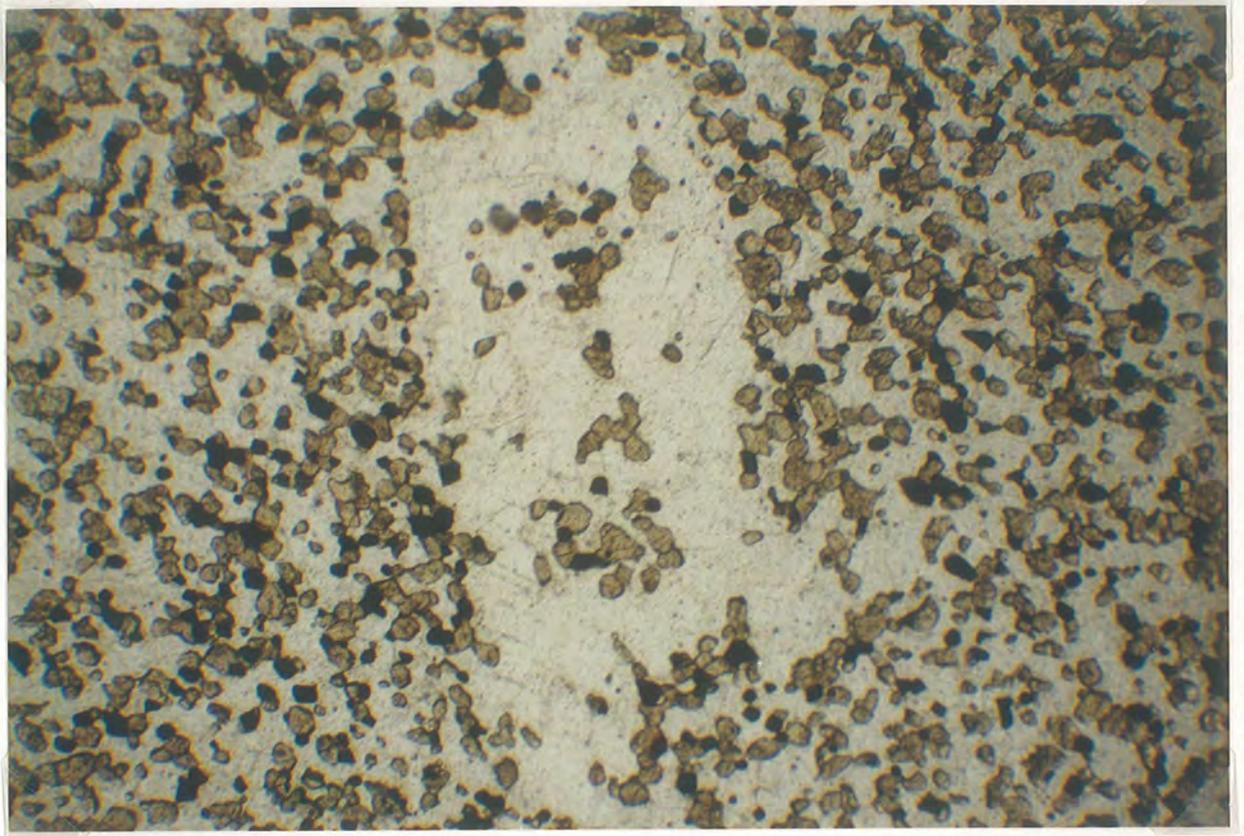
poorly constrained. They are, however, clearly older than an elongate body of coarse heterogeneous leucogabbro which occupies much of the valley of the Abhainn Chro Bheinn, because this has not been affected by high - grade metamorphism. As much as 80% of this later intrusion is made up of euhedral equant feldspar crystals. These lack the distinct calcic cores of equivalent phenocrysts in feldspar - rich rocks of the Hypersthene Gabbro itself. Interstitial to ophitic augite is the only pyroxene ( unlike the Hypersthene Gabbro ) and occurs with interstitial opaques and a significant amount of quartz - alkali feldspar granophyric intergrowth. This latter is unevenly distributed on a scale of a few centimetres or more. The intrusion is also heterogeneous on the outcrop scale, with decimetre to metre scale diffuse streaks and blebs of coarser, more feldspathic rock.

The other gabbros which outcrop around Glebe Hill are also unaffected by high - grade metamorphism but are petrographically distinct from this intrusion, which is referred to in this work as the Abhainn Chro Bheinn gabbro. This is especially true of the gabbros on the eastern side of the hill. Those at its southeastern end contain coarse to pegmatoid plagioclase - augite pods without quartz or opaques, in a host gabbro which lacks olivine and orthopyroxene and is opaque - poor. The outcrops on the eastern side of the valley to the east are finer grained and contain olivine. It seems likely that a number of small gabbroic intrusions are present around Glebe Hill, all of which postdate the high grade metamorphism seen in the probable Marginal Border Group rocks and in the earlier country rocks to the south. They also seem to postdate the intensely porphyritic dolerite cone sheets which themselves postdate the earlier of the two phases of metamorphism recorded by the basalts, picrites and early porphyritic dolerites ( see below ), although exposure in the relevant area on the eastern side of the hill is not good enough for this to be certain. Alteration at low hydrothermal metamorphic grades is, however, common in these rocks: the assemblage epidote + chlorite + quartz + calcite or sphene is typically present. This later alteration does not affect a number of felsite and feldspar - phyric basaltic dykes, striking  $080^{\circ}$ , which cut the Abhainn Chro Bheinn gabbro and resemble the Group 4 minor intrusions of the An Acairseid to Kilchoan sector.

The country rocks in the Glebe Hill area fall into four groups:

- 1). **Originally feldsparphyric basic rocks**, which outcrop on the northern and western side of the hill, close to the contact. These have been completely recrystallised, including the plagioclase phenocrysts, which form aggregates of feldspar grains with a preferred optic orientation ( Plate

Plate 3.63A and Plate 3.63B. Two views of a granular pseudomorph after a plagioclase porphyrocryst in granular basic hornfels at the northern end of Glebe Hill ( sample 50 ). Note the euhedral outline of the pseudomorph in 3.63A and the preferred optic orientation of the subgrains in 3.63B. 3.63B taken under crossed polarisers with  $1 - \lambda$  plate inserted. Both plates have a 1.5mm field of view.



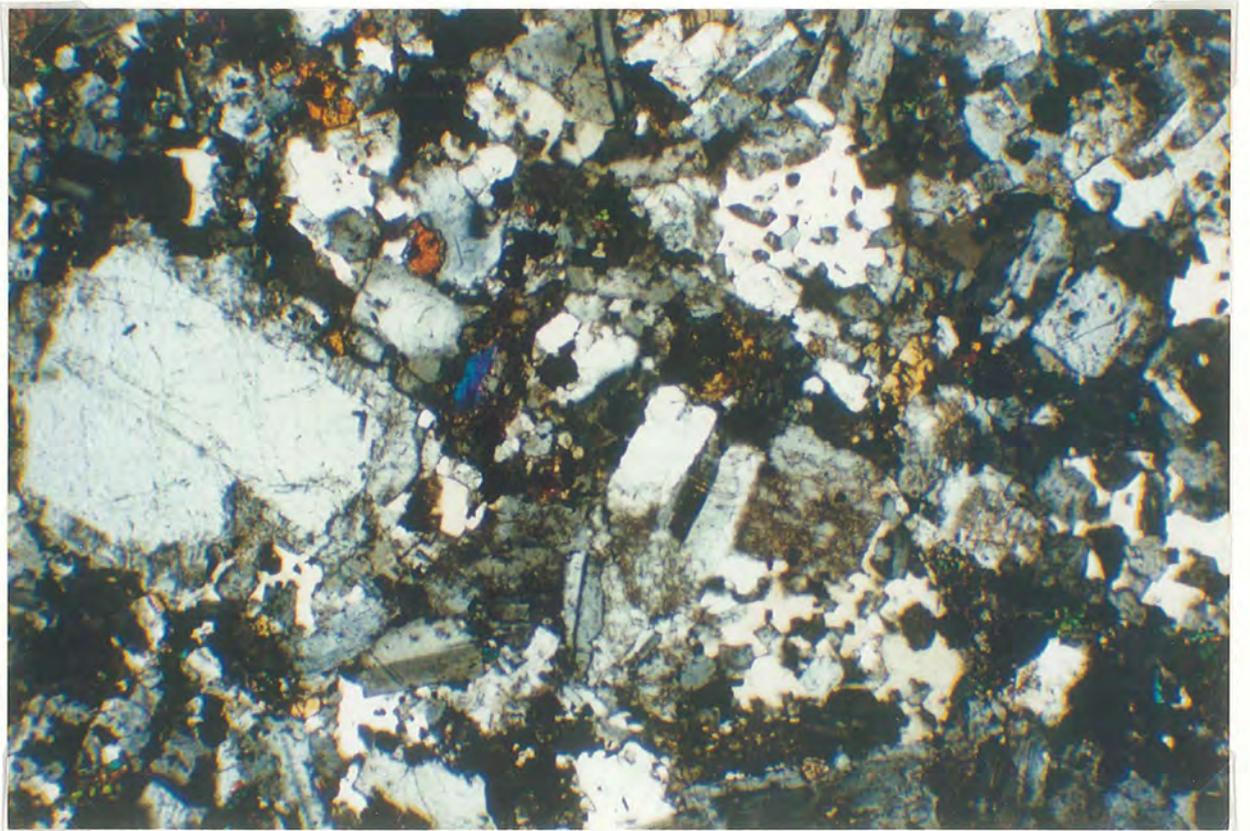
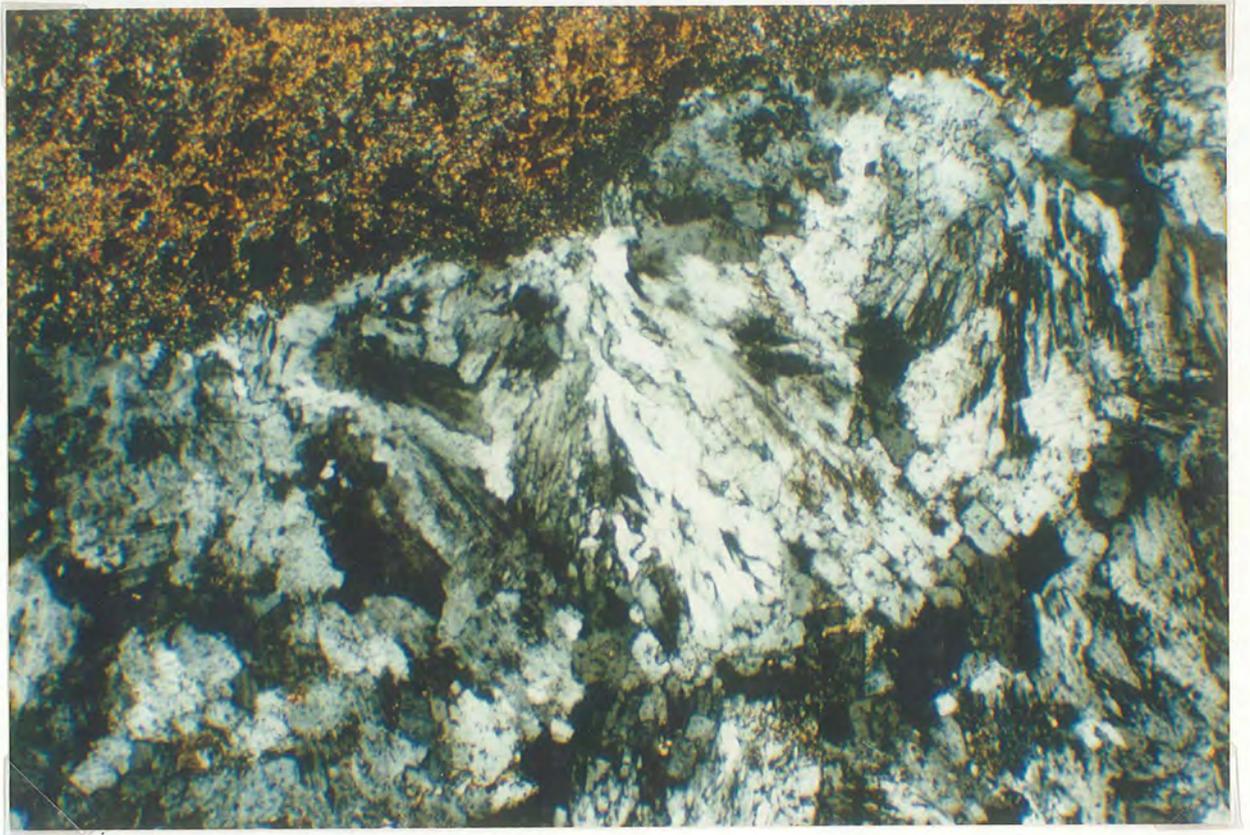
3.63A & B ), to granular, texturally equilibrated rocks with the assemblage calcic plagioclase + augite + olivine + opaques. This is a typical igneous assemblage and indicates that these rocks were not altered prior to metamorphism, in contrast to the adjacent basalts ( see below and Chapter 2 ). Although no convincing intrusive contacts were found, this suggests that these rocks were originally minor dolerite intrusions emplaced after alteration of the basalts. Post - high grade metamorphic alteration in these rocks only occurs around fractures, and is characterised by the formation of poikiloblastic red biotite.

2). **Amygdaloidal and non - amygdaloidal basalts and picrites.** These are also mainly granoblastic - textured with few recognisable relict igneous textures, except at the southern end of the hill. The absence of the latter may in part have been caused by pre - metamorphic, near - surface alteration of these rocks which also resulted in their anomalous compositions ( see Chapter 2 ) and high content of orthopyroxene. At the northern limit of their occurrence, within 5m of the contact at 48066478 ( the spinel hornfels locality discussed below ) these rocks are granoblastic basic hornfelses, with the assemblage plagioclase + orthopyroxene + olivine  $\pm$  augite + minor opaques and only indistinct olivine - orthopyroxene rich relicts after amygdales. Around the northern summit of the hill ( 480646 ), the metabasalts contain the silica - deficient assemblage plagioclase  $\pm$  orthopyroxene + olivine + opaques + augite + pargasitic hornblende and a variety of meta - amygdales, filled with olivine or orthopyroxene, or aegirine-augite  $\pm$  garnet  $\pm$  wollastonite. At the southern end of the hill, around 481643, relict plagioclase laths, set in a granular matrix, are common, and poikiloblastic opaques and pargasitic hornblende are major components of the rocks. The metaamygdales in these rocks are compositionally distinct, being filled with granular and sheaf - like clusters of feldspars ( Plate 3.64 ). The increasing abundance of the hornblende and opaques and the better preservation of igneous textures point to a gradual southward decrease in metamorphic grade along the hill ( see Chapter 5 for estimates of the conditions of this metamorphism ). The next outcrops of metabasaltic rocks to the south, at 48276380, contain a low grade, chlorite rich, alteration assemblage and well - preserved igneous textures. This alteration appears to be M0 in age. The high - grade metamorphic assemblages in the rocks on the hill are pristine, except around rare fractures, where olivine is replaced by talc + opaques, biotite is sometimes present, and opaque granules occur in pyroxenes. This suggests retrogressive hydration and oxidation associated with fluid flow along the fractures.

This group of rocks was found during the course of this work to enclose the hercynite - plagioclase -

**Plate 3.64.** Meta - amygdale in pargasitic hornblende rich altered - basalt hornfels. The amygdale is filled with granular and sheaf - like clusters of feldspar. Sample 254, from southern end of Glebe Hill. Crossed polarisers, width of field of view 14mm.

**Plate 3.65.** Porphyritic microgranodiorite, Glebe Hill ( sample 248 ). Note poikilitic quartz in groundmass at centre and at right of field of view. Crossed polarisers, width of field of view 3mm.



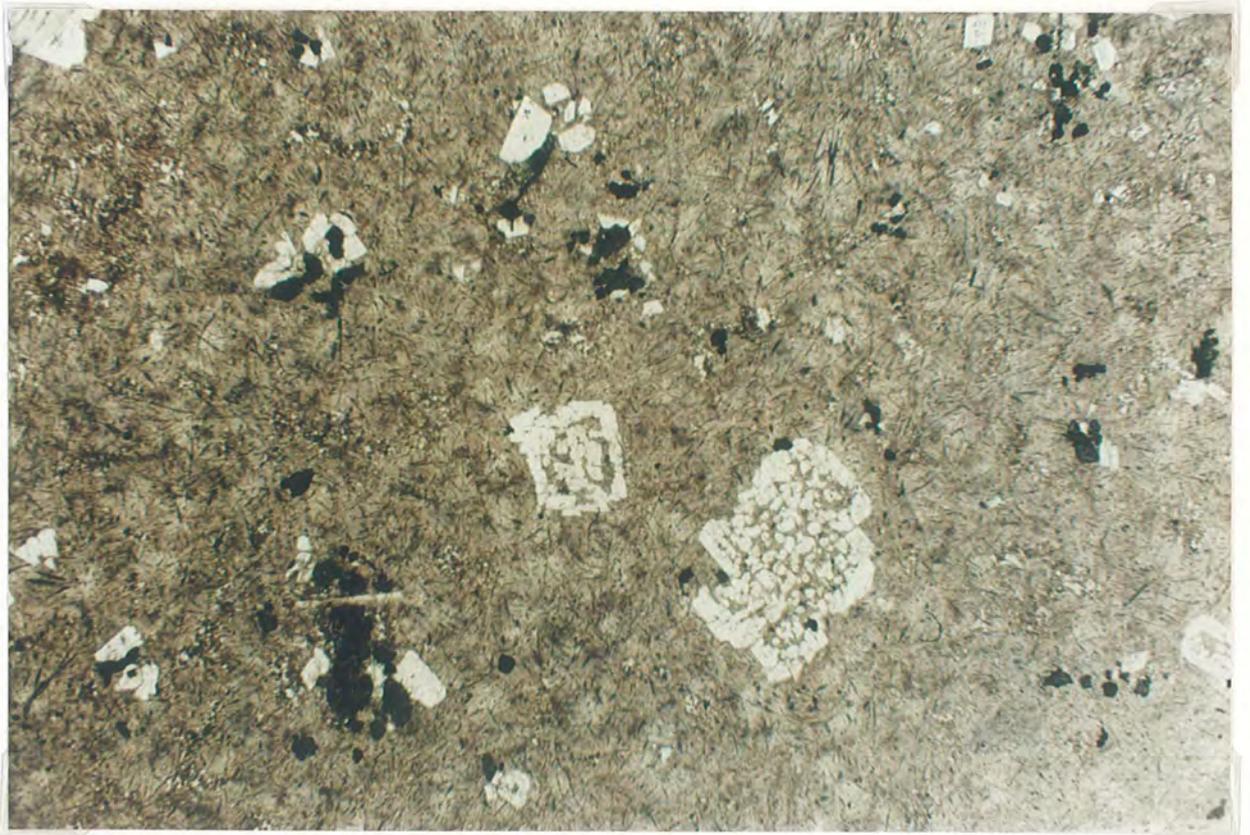
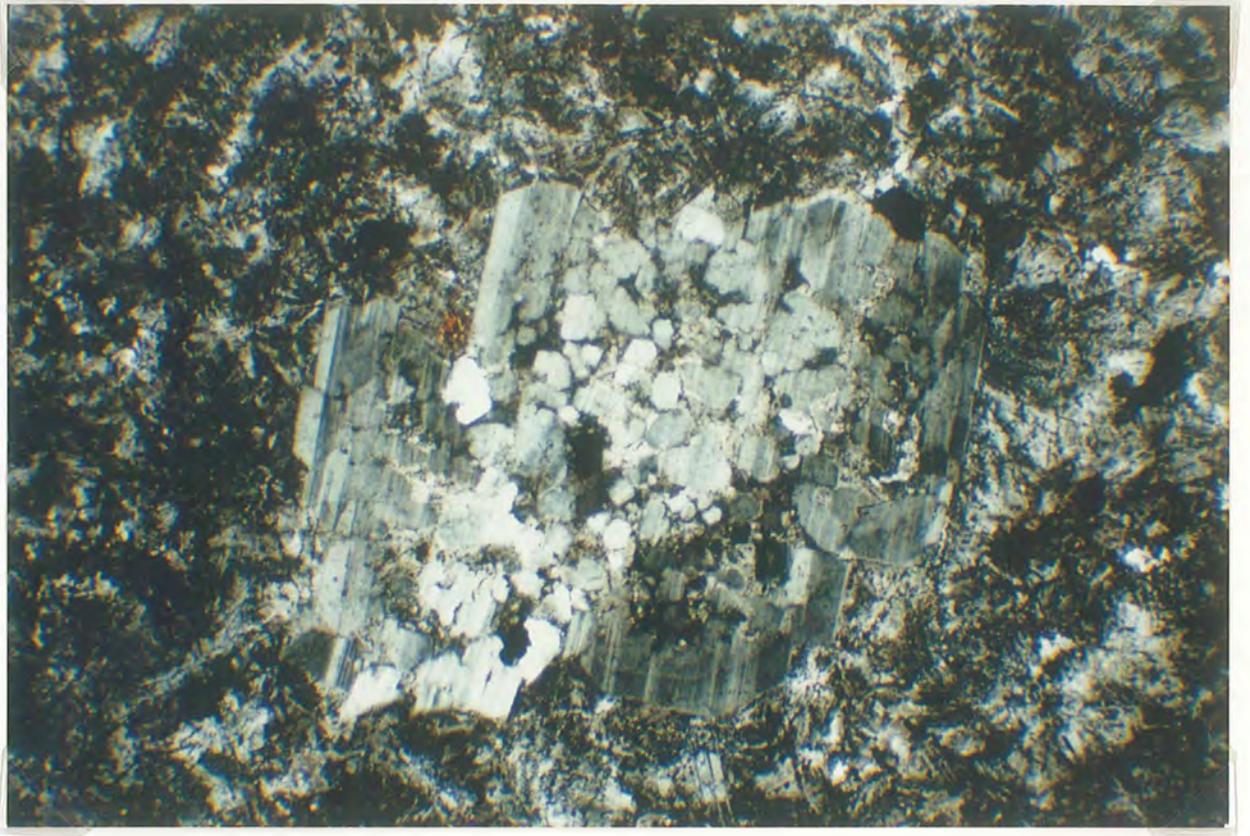
corundum hornfels first described by H.H.Thomas ( in Richey et al. 1930 ). Although the outcrops examined, at 48066478, are within 5 metres of outcrops of the Abhainn Chro Bheinn gabbro, they occur within granulitic - textured metamorphosed altered basalts ( see above ) and there is no evidence now visible to suggest that these are enclosed in rocks of the Hypersthene Gabbro, as was originally suggested by H.H.Thomas ( see section 2.4 ). The opaque phase in the spinel - rich hornfels is ilmenite and they lack magnetite. This implies a very low  $fO_2$  ( see Chapter 5 ) during metamorphism of these rocks, in contrast to the high  $fO_2$  of metamorphism in the magnetite - rich gabbros described above, which are believed to form part of the Hypersthene Gabbro. This implies either strong local variation of  $fO_2$ , or the existence of two separate phases of metamorphism, in which case the oxidative metamorphism in the gabbros would correspond to the retrogression observed along fractures in the metabasalts ( see above ).

**3). Post - highest - grade - metamorphism felsic intrusions.** These fall into two sub - groups ( see Fig. 3.19 ). One of these is a series of subvertical sheets, at most a few metres wide, on the west side of the hill between 47976442 and 48006438, with a strike of about  $060^\circ$  and associated with felsic veins in the host porphyritic basic hornfels. These are composed of porphyritic microgranite, with corroded andesine plagioclase phenocrysts, overgrown by euhedral plagioclase rims, set in a rather granular groundmass. The groundmass assemblage is quartz + alkali feldspar + augite + opaques ( Plate 3.65 ). Numerous aphyric basic hornfels inclusions are present, some of which are rounded and have crenulate chilled margins, but the rocks differ from those of the Sgurr nam Meann intrusion, which lies along strike to the west, in being more strongly hornfelsed. They appear to predate the second phase of metamorphism in the area ( see below ).

The second sub - group is formed by a single intrusion: a felsite sheet which occupies much of the summit region around 481645. This is a porphyritic rock with abundant large sieve- and negative sieve - textured corroded feldspar porphyrocrysts ( Plate 3.66 ). Negative sieve texture is here defined as that shown by clusters of rounded feldspars, with similar optic orientation but separated by melt channels, which define the euhedral or subhedral remains of a larger feldspar grain. The origin of this texture and its genetic relationship to normal sieve texture, which is formed by arrays of rounded melt pockets in a continuous crystal host ( the 'photographic negative' of negative sieve texture ) are discussed in Chapter 4. The remainder of the rock is made up of rare glomerocrysts of rounded and corroded feldspar; augite ( largely replaced by granular opaques ); even rarer elongate

**Plate 3.66.** Plagioclase porphyrocryst with negative sieve - textured core in felsite, Glebe Hill ( sample 250/1 ). Note straining and subgrain development in rim of porphyrocryst: this probably predates quenching of the parent magma as the groundmass is undeformed. Crossed polarisers, width of field of view 3mm.

**Plate 3.67.** Negative sieve - textured plagioclase porphyrocrysts and plagioclase - augite - opaques glomerocrysts ( microxenoliths ? ) in sample 250/1. Plane - polarised light, width of field of view 14mm.



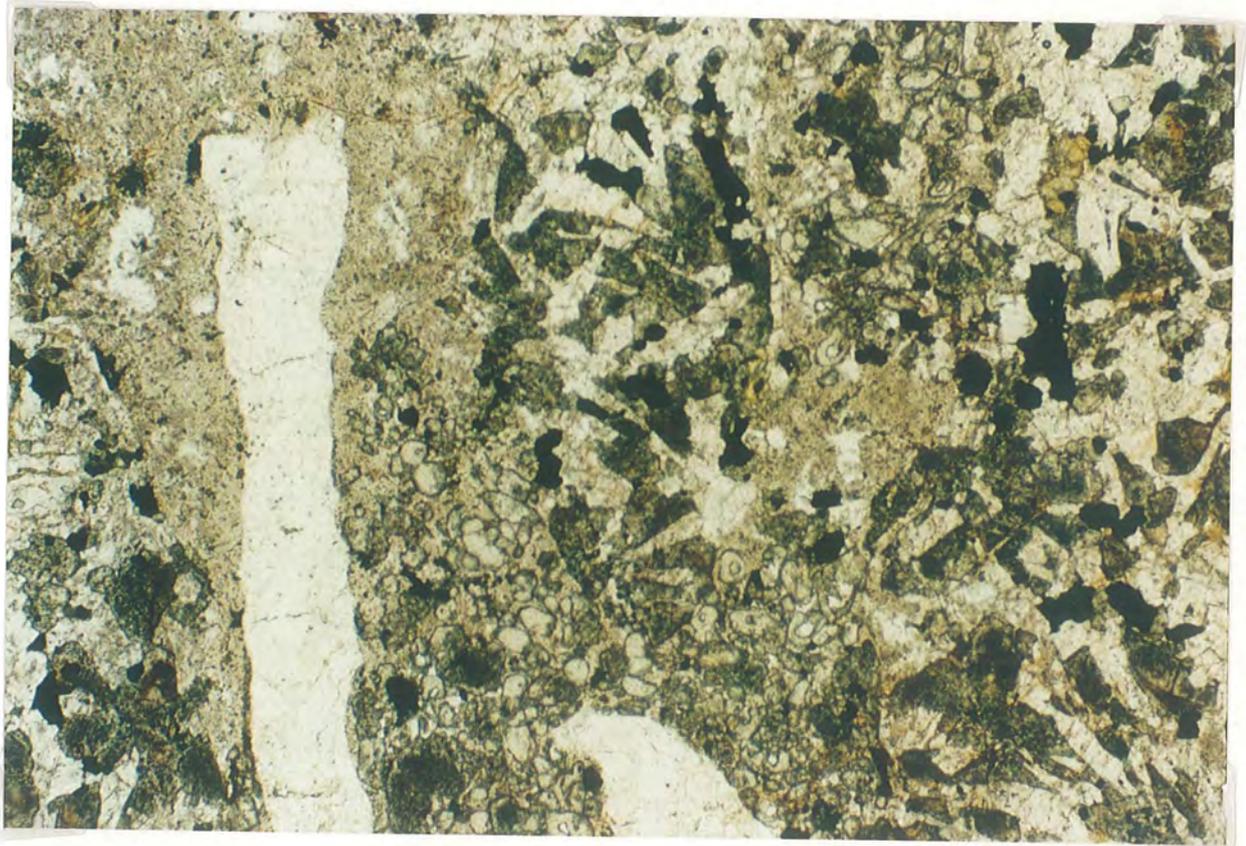
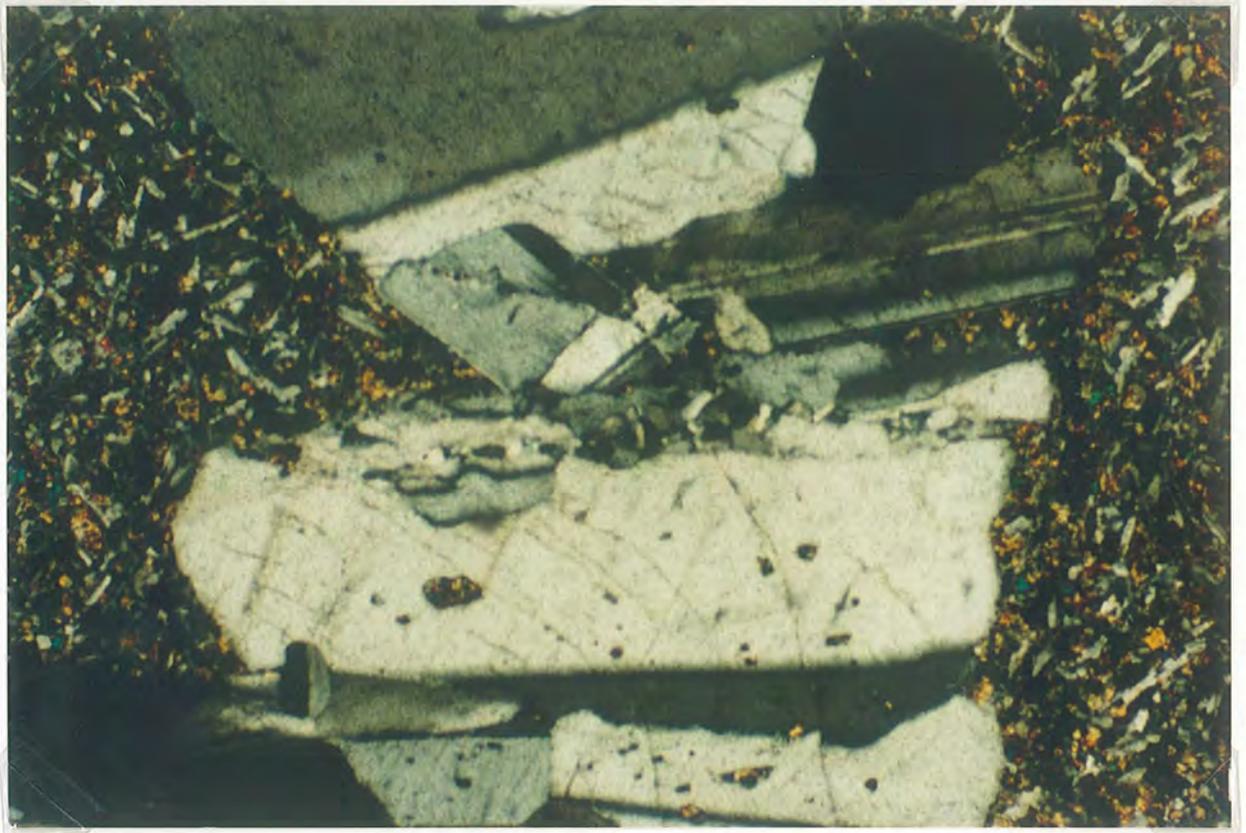
( possibly quench ? ) augites, also partly replaced by opaques; and a devitrified and hornfelsed microgranophyric - textured groundmass, with prominent opaque crystallites and patchy replacement by granular quartz - chlorite - alkali feldspar aggregates of probable hydrothermal origins, that forms more than 80% of the rock ( Plate 3.67 ). The low crystal content of the rock, and in particular the very low abundance of magmatic phenocrysts, make it comparable with the M2 felsites found on the northern margin of the Hypersthene Gabbro, rather than with M1 microgranites: the same is probably true of the microgranite sheets on the eastern side of the hill. This has the corollary that although the felsite itself has been quenched in the manner to be described in Chapter 5, this does not imply a quenching event in its host rocks.

4). The final group of intrusions on Glebe Hill are a series of **porphyritic and microxenolithic dolerites**, intruded as cone sheets with shallow northerly dips ( 20 - 40° ). The porphyrocrysts are all plagioclase feldspar and the microxenoliths are anorthositic. The single crystals are rounded and sieve textured, whilst the microxenoliths are identifiable as such ( as opposed to magmatic glomerocrysts ) because of the presence of strained, sutured and granular ( i.e. texturally re-equilibrated ) grain boundaries between the crystals in the microxenoliths ( Plate 3.68 ). The groundmass of these dolerites is commonly rather granular but retains relicts of an original micro - poikilitic texture, with plagioclase laths set in subophitic augite which has partially recrystallized to granules. Subsequent alteration mainly takes the form of intense oxidation of the augite, which is crowded with opaque inclusions; minor secondary phases include biotite, even rarer haematite, and small amounts of later chlorite and green amphibole. These rocks therefore predate a phase of moderate to high grade metamorphism at high  $fO_2$ , analogous to that seen in the granular gabbros at the northern end of the hill.

The same phase of metamorphism, which is characterised by oxidation without complete recrystallisation, is also recorded by a near - aphyric dolerite hornfels sheet at 48146436, within a few metres of the heterogenous gabbros which outcrop at the southeastern end of the hill. This dolerite, which was not affected by the intense recrystallisation and associated pargasitic hornblende formation developed in the adjacent basalts, contains a pre - metamorphic quartz vein which has undergone partial melting, by reaction with the host basalt, to form a clinopyroxene - phyrlic melt which has subsequently been quenched to a variolitic quartz - feldspar intergrowth ( Plate 3.69 ). This implies that the oxidative metamorphism in this area ended in a quenching event. However, this cannot

**Plate 3.68.** Laminated anorthositic microxenolith cut by annealed microfault, in post - high - grade - metamorphic porphyritic dolerite cone sheet, Glebe Hill ( sample 253/3 ). Crossed polarisers, width of field of view 3.5mm.

**Plate 3.69.** Fused quartz - rich hydrothermal veins in post - M1 dolerite, southern end of Glebe Hill. Note that the veins are heterogenous, containing felsitic and diopside - rich segments ( corresponding to carbonate - rich vein segments ? ) as well as relict quartzitic inclusions. Plane - polarised light, width of field of view 3.5mm.



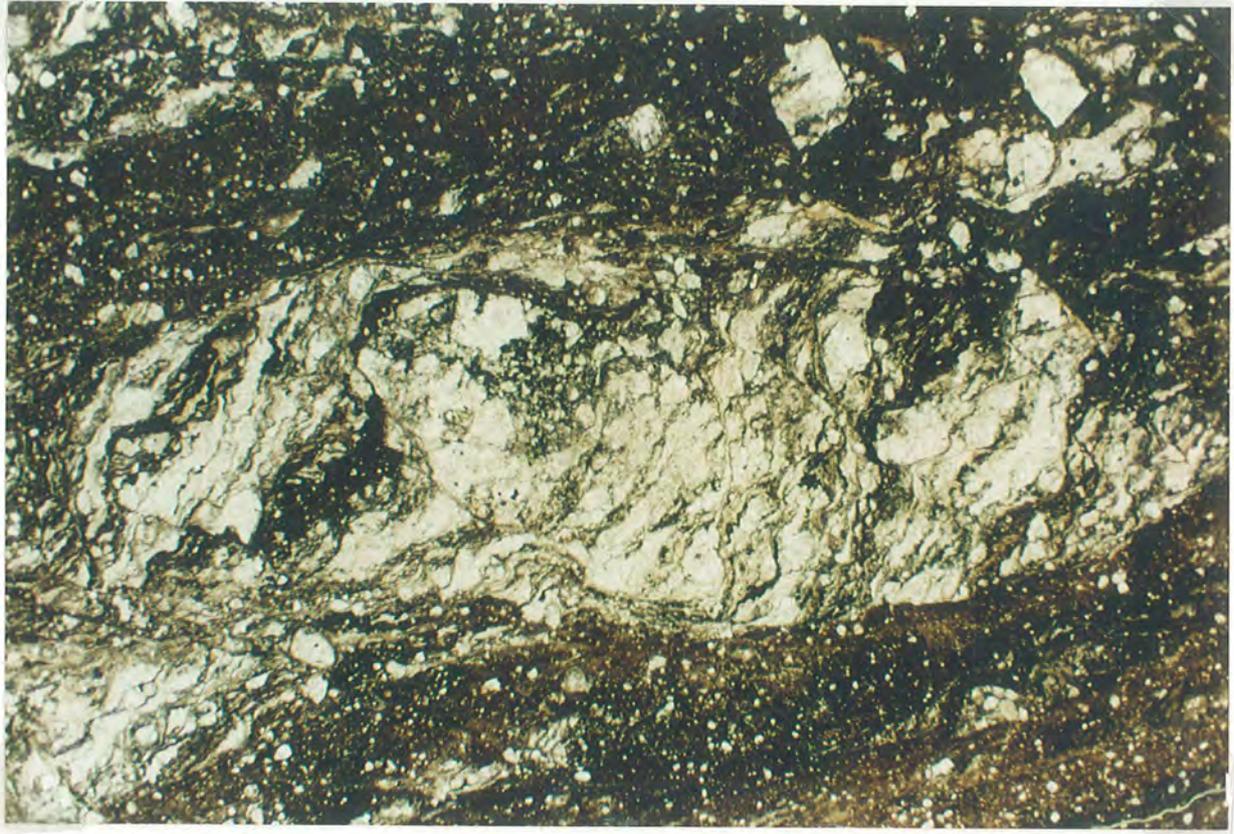
be correlated with the end - M1 quenching event observed around the contact zone in other sectors of the Hypersthene Gabbro and its contact aureole, because the preceding metamorphism affects rocks which appear to be part of the MBG ( the magnetite - rich gabbros and heterogenous rocks around the Abhainn Chro Bheinn described above ). A further implication of the melting of this quartz vein is that the oxidative metamorphism and the metamorphism developed in the basalts must have been separated by a period of hydrothermal fluid circulation, in which the vein formed: this is analogous to the sequence of events in the metamorphosed cone sheets on the western side of Hill 90, to the south of Glendrian Bay ( section 3.2.2 ).

The timing and incomplete nature of the melting event recorded by the quartz vein suggests that it was caused by the immediately adjacent small gabbro intrusions, which are themselves unaffected by the oxidative metamorphism. The small size of the heat sources would also explain why the felsite intrusion on the top of the hill ( some 150m to the north ), which is cross - cut by a member of the strongly porphyritic dolerite suite, shows no sign of such high grade metamorphism. On the other hand, the oxidative metamorphism in the basic rocks is developed all along the hill, suggesting that the local thermal anomaly which melted the quartz vein was superimposed on a much broader region of more moderate - grade metamorphism.

The history of metamorphism in the Glebe Hill area can therefore be summarised as follows:

- 1). Early high to very high grade metamorphism, which increases in grade and intensity towards the north and the contact with rocks probably belonging to the Marginal Border Group of the Hypersthene Gabbro, but does not affect the latter. It is therefore equivalent to M1 elsewhere in the contact aureole of the pluton.
- 2). Cooling to low temperatures, and formation of quartz veins. The microgranodiorite and porphyritic felsite intrusions and the xenocrystic/xenolithic dolerites were emplaced during this period.
- 3). Moderately high grade oxidative metamorphism. Textures and mineral assemblages formed during this are similar to those developed at the highest grades of M2 metamorphism in the An Acairseid - Kilchoan sector, except that biotite is absent, possibly indicating slightly higher grade conditions in these rocks. The melting of the quartz vein at the south end of the hill may have

**Plate 3.70.** Cataclastic rock from the Glas Eilean fault zone. The plate shows a deformed clast, about 1.5 centimetres long, in an ultracataclastic fault within the fault zone. The clast itself is cut by earlier cataclastic or pseudotachylitic veins, some of which have been annealed and all of which are deformed to varying extents. This indicates a long history of movement even on this particular fault within the fault zone. Sample 359A. Plane - polarised light, width of field of view 14mm.



occured at the end of, or just after, this phase of metamorphism.

4). Low grade hydrothermal metamorphism, which affects all the rocks in the area, including the Abhainn Chro Bheinn gabbro and the other small gabbro intrusions.

If the correlation of the oxidative metamorphism with early M2 metamorphism elsewhere around the pluton is accepted, it follows that M1 and the peak of M2 in this area were separated by a period of lower grade hydrothermal metamorphism, in which the felsites were quenched and the quartz vein probably formed, as seems to be the case in places on the northern margin of the pluton ( sections 3.2.2 and 3.2.3 ). It also follows that the geochemistry of the xenocrystic/ xenolithic dolerites should be similar to that of other early M2 minor intrusions, whilst the Abhainn Chro Bheinn and other later gabbros in this area should be comparable to late- or post- M2 intrusions, such as the Group 3 cone sheets of section 3.2.5 or one of the various suites of rocks in the Inner Series. These predictions are tested in Chapters 4 and 6.

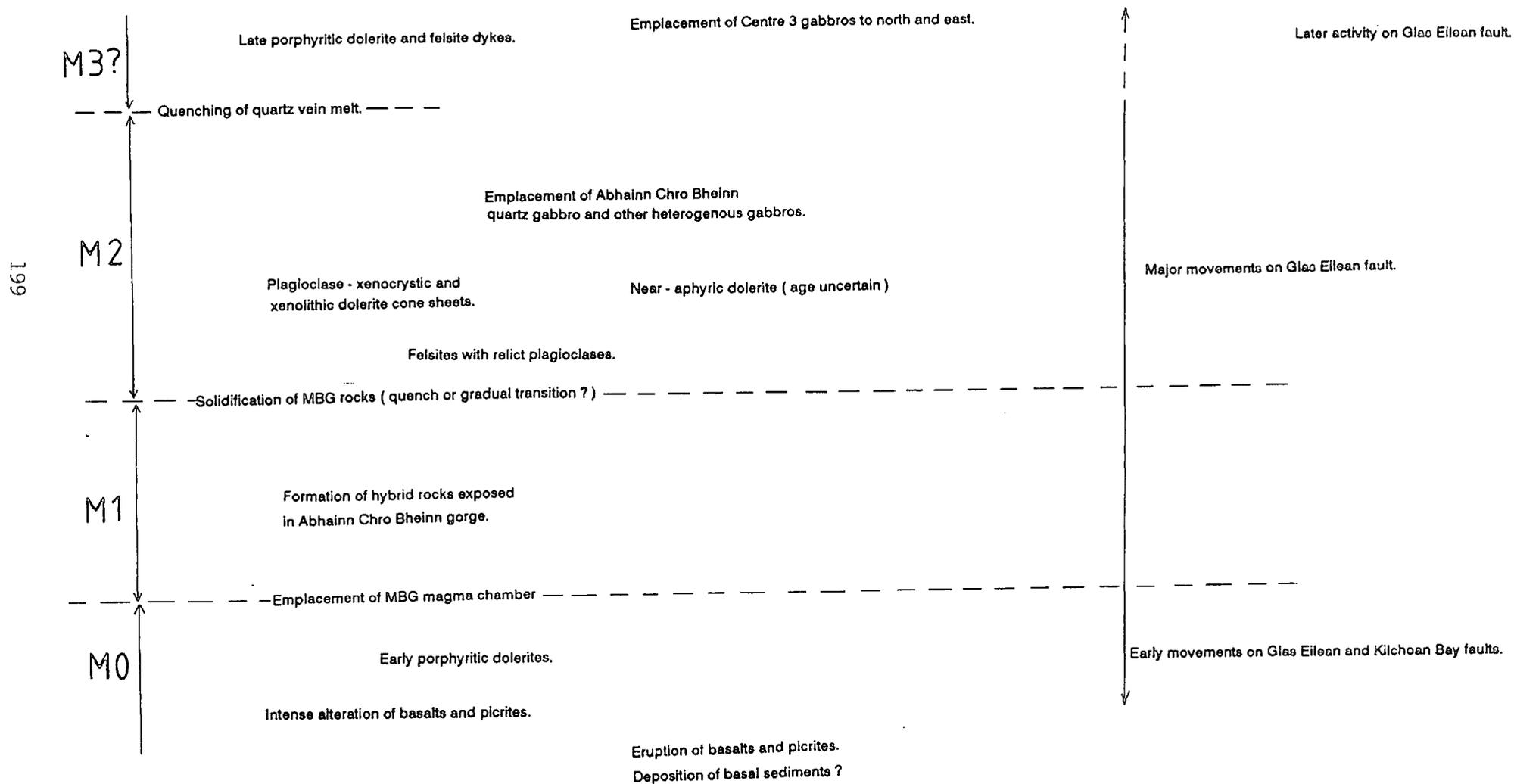
Further constraints on the chronology of the Glebe Hill area are provided by the age relationships of the faults which bound the outcrop of the basalts, particularly those of the Glas Eilean fault. This is exposed on Glas Eilean itself and on a promontory to the north, around 485630. Here it is a zone 1.5 to 3 metres wide, striking  $026^{\circ}$  and dipping northwest at  $70 - 80^{\circ}$ , of narrow anastomosing cataclastic fault zones. These are at most a few centimetres wide and enclose lenses, up to 5m long but less than 0.6m wide, of less strongly deformed Moinian metasediments and dolerite, probably derived from adjacent cone sheets. The fault zones themselves are composed of intensely deformed cataclastic rocks which show evidence for polyphase deformation in the shape of earlier annealed crush zones which occur within rotated clasts in the finer - grained, un - recrystallised, later cataclasites ( Plate 3.70 ). The clasts and the recrystallised cataclastic rocks contain low grade metamorphic chlorite + epidote  $\pm$  actinolite assemblages which are especially well developed in deformed metabasic rocks. The presence of this assemblage, coupled with the fact that the fault cuts all the cone sheets which it intersects on and around Glas Eilean, suggests that it was active well into the period of M2, although as noted in 3.2.5.5, the age relationships of cone sheets outside the high - grade M1 aureole to M2 are rather uncertain. As was shown in Chapter 2, this fault, or possibly the Kilchoan Bay fault to the west, must have been active as a westward - downthrowing structure at least as early as the formation of the unconformity at the base of the Tertiary basalts, but the width of, and the

intense deformation within, the post - cone sheet fault zone implies substantial post - cone sheet movements. Later movements along the general line of the Glas Eilean fault are also indicated by the presence of diffuse granular anastomosing shear zones cutting the later microgabbros and gabbros to the east of Glebe Hill, around 483647 ( see Fig. 3.19 ). These do not show intense hydrothermal alteration of the primary pyroxenes and olivines caught up in them, and seem to have formed soon after emplacement of these rocks, when they were still at near - magmatic temperatures. Although striking parallel to the expected line of the main fault these shear zones dip steeply eastward and cannot be the main fault for geometrical reasons: the occurrence of basaltic hornfelses as far east as 48286475 suggests that the main fault should lie further to the east. Mapping by Richey et al. ( 1930 ) suggests that minor movements along the line of the Glas Eilean fault may have continued until after the emplacement of the youngest intrusions in 'Centre 3' of the intrusive complex ( Fig. 1.2 ).

The age relationships of the fault along the northwestern side of the Glebe Hill basalts, the Kilchoan Bay fault, were not constrained during the course of this work because no exposures of it were found. The maps of Richey et al. indicate that movement on it must have ended at an early stage, because it is truncated by the Marginal Border Group, albeit in a poorly exposed area. Its age relationships would then be similar to those of the similarly outward - downthrowing faults between the Hypersthene Gabbro and the northward extension of the Glas Eilean fault to the east of Glendrian Bay ( section 3.2.2 ).

In conclusion, age relationships in the Glebe Hill area are complicated by the later small gabbroic intrusions and by the later intrusions of 'Centre 3' to the north. This means in particular that the age of the late, low grade ( M3 ? ) hydrothermal metamorphism recorded in the small gabbro intrusions as well as in the older rocks is very uncertain. Fig. 3.20 shows age relationships in the area in the usual form. It is constructed in the acceptance of the evidence presented above for the correlation of the early, very high - grade metamorphism with M1 and of the high  $fO_2$  metamorphism with M2 as they are defined elsewhere around the contact. The main difference in the sequence of metamorphism is then the localised high grade metamorphism around the small gabbro intrusions and the subsequent quenching event. The transition from M1 to M2 may also have been accomplished in a rapid cooling event which is not directly recorded outside the Hypersthene Gabbro in this area because of the absence of fusible lithologies, nor inside it, possibly because of

Fig. 3.20. Age relationships in the Glebe Hill area.



later overprinting during M2. It would, however, account for the extremely well - preserved state of the high grade assemblages in rocks affected by M1.

**3.2.7. A summary of the rocks and age relationships, up to the middle of M2, observed in the Contact Zone of the Marginal Border Group and the adjacent country rocks.**

Many of the later rocks, structures and metamorphic events present in the MBG and the adjacent country rocks are coeval with the development of the Inner Series of the pluton. A synthesis of the overall history of the MBG is therefore deferred to section 3.5, whilst the scope of this section is restricted to the unusual and complex structure and petrology of the contact zone suite of the Marginal Border Group and its evolution during M1, the quenching event at the end of M1, and the earlier part of M2.

The contact zone suite is developed at the steeply outward - dipping outer margin of the Marginal Border Group. It formed against the outer wall of the large, mainly basaltic magma chamber which the MBG represents. The contact mainly dips outwards at 60° or more and is dissected by near - vertical radial faults, faults trending 140 - 150° which form part of the regional fault system, and, most importantly, by a set of roughly concentric inward - dipping, inward - downthrowing extensional faults which give the contact a saw - toothed profile in radial cross - sections. These concentric faults were mainly active during high grade ( M1 ) metamorphism, at which time they rooted out into the molten magma chamber, but some of them continued to be active during M2.

The scale and degree of development of the Contact Zone suite varies greatly around the pluton. At many localities, fine - grained homogenous microgabbros occur within a metre of the contact and the only felsic rocks present are rare thin microgranitic veins. At the other extreme, the contact zone at Glendrian Bay, Duin Bhain, Ardnamurchan Point, Hill 210 and northeast of Glebe Hill is a zone, 50 - 100m wide, of very varied rocks with a high proportion of felsic and heterogenous hybrid material. In addition, a number of rheomorphic breccia intrusions of M1 age occur outside the pluton.

The main outer contact between the MBG and the country rocks is invariably sharp, although it is sometimes lobate or irregular because of mesoscopic rheomorphism in the country rocks. There is little field or petrographic evidence to suggest that large - scale mixing of country rock melts and

MBG magmas took place anywhere other than in and around the downfaulted blocks or pseudo-screens which projected into the magma chamber and which were produced by M1 movements on the concentric extensional faults. The process appears from field evidence to have involved formation of high - degree partial melts in the downfaulted blocks, followed by mobilisation of those melts and magma - mixing between the anatectic magmas and more mafic magmas. It follows that the formation of anatectic magmas, and the contamination of mafic magmas in the Marginal Border Group magma chamber by those magmas, was at least in part triggered, and hence controlled by, the expansion of the magma chamber and consequent deformation of the chamber walls. The significance of this control for the geochemical evolution of the magmas in the magma chamber is examined in Chapter 4.

Wall rock xenoliths within the contact zone mostly occur in swarms each of which is dominated by a particular refractory xenolith lithology. The dominant lithology varies between swarms but is typically metasedimentary, although mafic igneous hornfels form minor components of some swarms. The swarms may occur in otherwise homogenous basic rocks but are often enclosed in hybrid rocks rich in a felsic melt component. Rocks transitional to the xenolith swarms are probably not present at the main contact: the most nearly comparable rocks are the refractory cognate blocks in rheomorphic breccias within the downfaulted block on Hill 210 and the pseudoscreen at Duin Bhain.

Contacts between magmatic rocks of M1 age in the contact zone are almost all various forms of liquid - liquid contacts ( the exceptions are the edges of the fragments of shattered basic pillows ). These indicate that during M1 these rocks occurred in situ as magmas of varying temperatures and compositions, which resulted in chilling of some magmas against others ( see Chapter 4 ). The M1 rocks present include quartz- and olivine- dolerites and microgabbros, ferrogabbros and diorites ( at a few localities only ), hybrids formed in single- and multi-stage mixing events, microgranodiorites and a variety of granitoid rocks with unusual mineralogies. These include diopside granite, hypersthene - cordierite granite and augite diorite. These granitoid rocks, and the microgranodiorites, occur intermittently, in discrete bodies each of which has a distinctive composition.

All of the contact zone rocks show, to varying extents, signs of rapid solidification under super-cooled conditions at the end of M1. This was associated with the formation of a network of closely

spaced tensile fractures. These fractures are commonly plugged by hydrothermal mineral assemblages, indicating that they acted as fluid flow conduits after their formation. Quench textures in these rocks range from radiating clusters and sheaves of plagioclase laths in dolerites to fine - scale microgranophyric intergrowths, quench crystals and variolitic to spherulitic devitrified glasses in the more evolved rocks, especially those anatectic rocks which were largely molten during M1. There is some spatial variation in the development of quenching around the pluton: devitrified glasses are much more common on the northern margin of the pluton, from Duin Bhain to Glendrian Bay, than elsewhere. However, microgranophyric and micropoikilitic textures which may have been produced by more advanced alteration of primary glasses are common in other sectors of the contact, so the absence of glasses from large parts of the contact zone may be the result of greater temperatures and durations of alteration in these areas rather than lower rates of cooling during quenching at the end of M1. This interpretation is supported by the greater intensity and higher peak grade of M2 metamorphism in these areas.

Whilst the quenching event in the country rocks and in the microgranodiorites at the outer edge of the contact zone appears to have been very rapid, the contact zone contains a number of rocks and structures which appear to have formed after its consolidation but before it crystallised completely. The most abundant of these are the fractionated, opaque - rich ferrogabbros and ferrodiorites which occur as irregular, sometimes pegmatoid pods and veins. The abundance of these is greatest in a distinct zone a few tens of metres inside the contact, which is particularly well developed at Ardnamurchan Point and Chorrachadh Mhor. A number of basic minor intrusions were emplaced in the still - hot contact zone during this period, mainly unchilled cone sheets but also the rooted dykes at Glendrian Bay. Finally, unchilled granodioritic and other felsic veins also formed within the Marginal Border group at this time.

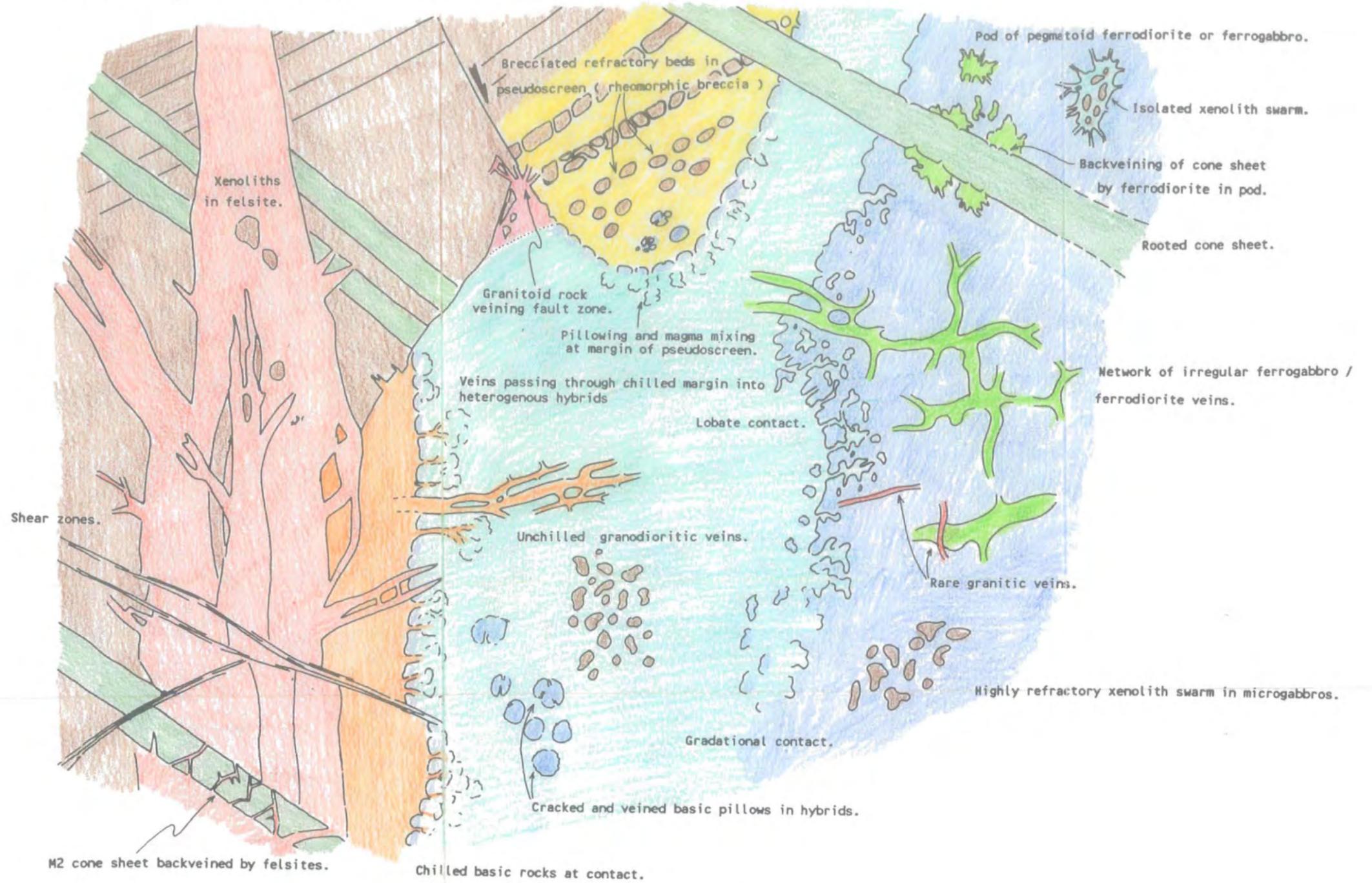
Post - quench, M2 minor intrusions are also present in the contact zone and the adjacent country rocks. Those most closely associated with the contact zone are the porphyritic felsites which occur in a relatively large multiple intrusion near Duin Bhain, a large sill - like body on Glebe Hill, and in numerous inward - dipping lensoid sheets and veins. Like the rocks of the contact zone and anatectic rocks in the contact aureole, these show evidence of quenching associated with formation of a tensile fracture network plugged by hydrothermal minerals. The groundmasses of these felsites are completely homogenous apart from the effects of devitrification ( see sections 4.2.2 and 5.3.2 ).

There are no relict glass shards or other structures in them which might suggest that the felsites were emplaced as tuffisites or 'intrusive ignimbrites', as has been proposed for the similarly fine-grained intrusive Loch Ba felsite in the Mull central complex to the south of Ardnamurchan ( Sparks 1988 ).

In addition to the felsic intrusions, a very large proportion of the dolerite cone sheets around the intrusion appear to have formed soon after the end of M1. As noted in section 3.2.2, this has the very important implication for cone sheet emplacement mechanisms that they formed at the steep sides of the pluton, rather than above an earlier buried intrusion.

Fig. 3.21 is a composite diagram which attempts to show all the rocks and structures developed in the contact zone and their mutual relationships in a single cross - section. These are by no means all developed in any one area of the contact: the outcrops around Duin Bhain probably come closest to showing all the features shown in this diagram.

Fig. 3.21. Generalised diagrammatic cross - section showing cross - cutting and contact relationships in the Contact Zone of the Marginal Border Group.



- |  |  |   |   |
|--|--|---|---|
|  | M2 Felsites.                               |  | M1 microgranodiorite.                           |
|  | Dolerite cone sheets.                      |  | Microgranitoid rocks of anomalous compositions. |
|  | Homogenous MBG microgabbros and dolerites. |  | Largely melted country rocks in pseudoscreen.   |
|  | Heterogenous hybrid mottled rocks.         |  | Country rocks.                                  |

### 3.3 The geometry and origin of the contact between the Marginal Border Group and the Inner Series.

This section is primarily concerned with the contacts observed between rocks assigned to the Marginal Border Group and those assigned to the Inner Series. However, because previous workers have considered that there are no major discontinuities within the Hypersthene Gabbro and that the transitions from one rock suite to another within it are gradational ( Richey et al. 1930; Wells 1954, 1978; Skelhorn & Elwell 1971 ) it is appropriate to first consider the features used, during mapping carried out as part of the present work, to identify rocks as belonging to the MBG or to the Inner Series ( IS ). These identifications were based on field, petrographic and mineralogical observations. Geochemical differences between the two groups of rocks are discussed in Chapters 4 and 6.

#### 3.3.1. Field, petrographic and mineralogical criteria for distinguishing rocks of the Marginal Border Group from those of the Inner Series.

The Marginal Border Group can be defined as those rocks associated with the highly distinctive contact zone suite of granitoid, hybrid and pillowed heterogenous rocks, pseudoscreens and xenolith swarms which was summarised in section 3.2.7. It includes the contact zone suite itself but also the homogenous, isotropic and mainly fine - grained quartz- or olivine- bearing dolerites, microgabbros and gabbros which have liquid - liquid, pillowed or lobate contacts with the contact zone suite and were therefore coeval with it, and various minor dolerite intrusions with unchilled margins against older rocks of the Marginal Border Group. These two latter groups are characteristically aphyric or sparsely plagioclase - phyric, contain orthopyroxene as a ubiquitous but minor phase and show moderate to low grade hydrous M2 alteration. Characteristic metamorphic phases include biotite, talc, opaques, hornblende and clinopyroxene. With very few exceptions, mainly in the Glebe Hill area, they do not show high - temperature alteration comparable to that which is near - ubiquitous in the Inner Series. Where present, the feldspar phenocrysts in the basic members of the MBG are relatively small, uncommon, weakly zoned and typically have labradoritic cores. A very small number of intensely zoned microphenocrysts with bytownitic cores do occur in MBG basic rocks but these are much smaller and much less common than the bytownite porphyrocrysts which are a prominent feature of the Inner Series gabbro-norites ( see below ). A variety of xenocrystic calcic feldspars and anorthositic microxenoliths may be present in MBG rocks, however.

The Inner Series essentially consists of a single major lithology and a number of very varied minor

ones. These are described and discussed in detail in section 3.4 and Chapter 6 but are introduced here to indicate the great variety of differences between the MBG and the IS. The dominant IS lithology is an olivine gabbronorite, near - troctolitic in a few outcrops, with a high modal content ( 10 - 20% ) of orthopyroxene and usually more than 5% modal olivine. Texturally, the gabbronorites vary from coarse ( grain sizes mainly 3 - 10mm ( feldspar ) and up to 5cm ( ophitic pyroxenes and olivines )) gabbroic - textured varieties to coarse doleritic rocks with large ( up to 3cm across ) ophitic pyroxenes. Ophitic orthopyroxenes are particularly well developed in the Inner Series dolerites and make them very distinct from the MBG dolerites and microgabbros. All contain large ( up to 2cm long ) feldspars with one or more euhedral to subhedral core zones. Microprobe analysis ( Chapter 6 ) of these crystals shows that they typically have calcic bytownite inner cores (  $An_{81}$  to  $An_{85}$  ), bytownite outer cores (  $An_{77}$  to  $An_{80}$  ) and zoned labradoritic rims. The two core zones are unzoned in most crystals. These crystals give most of the gabbronorites a porphyritic texture but are not immediately obvious in the coarsest varieties. Small anorthositic gabbro and anorthosite xenoliths are common in these rocks but are finer grained than the calcic plagioclase porphyrocrysts. Although most of the gabbronorites appear isotropic at outcrop, isolated occurrences of lamination are visible in the field and a weak plagioclase lamination is usually apparent in thin sections of the coarser gabbroic textured varieties, in contrast to the completely isotropic MBG rocks. A final characteristic feature of the gabbronorites is the presence of indistinct but relatively sharp planar shallowly-dipping internal contacts. These occur between bands of rock up to tens of metres thick which differ slightly in composition and/or texture. The origin of these internal contacts is discussed further in section 3.4.1; for present purposes their significance is that they never occur within the MBG.

The various rock types associated with these gabbronorites probably make up no more than 5% of the Inner Series but are extremely diverse. They include both rocks older and younger than the gabbronorites:

- 1). **Granular xenoliths**, with sharp contacts against the host gabbronorites or laminated rocks ( of Group 2 below ). These contain various mineral assemblages ( see section 3.4.2 ) but most contain compositional banding or layering and appear to be layered igneous rocks: similar xenoliths are very rare in the MBG, whilst only one xenolith with a probable sedimentary protolith was found in the IS, and is itself enclosed in another xenolith.

2). **Laminated and Layered Rocks.** These are mainly aphyric, or sparsely feldspar phyric, and rather fine-grained and granular gabbros and gabbronorites with well-developed plagioclase lamination and weakly developed modal layering. A few granular anorthosite bands and pods also occur within these rocks. They occur in well-defined areas which in at least most cases can be shown to be screens between younger gabbronorites which have mainly shallowly inward dipping contacts against the laminated rocks.

3). **Coarse Anorthositic Gabbros and Anorthosites.** These also occur in screens and xenoliths enclosed in the gabbronorites, and are characterised by a lack of orthopyroxene. Olivine is less abundant in most of these rocks than it is in the gabbronorites.

4). **Pegmatoid pods and veins.** Many of the gabbronorites contain very coarse, pegmatoid-textured pods and veins, with diffuse and irregular margins, of augite rich gabbro. A few occurrences of plagioclase - rich coarse veins and pods were also found. Some of the gabbroic pegmatoids contain mafic granophyre cores. These rocks are different from the pegmatoid veins in the MBG in being coarser grained still and in having a lower content of opaque phases. Apart from the granophyre cores of these bodies, only one occurrence was found of coarse unchilled granitoid veins which do not appear to be associated with later intrusions.

5). **Secondary Pyroxenites.** Irregular, non-displacive tabular bodies and veins of coarse pyroxenite, olivine pyroxenite and pyroxene-rich gabbro which crosscut lamination and intrusive contacts occur in some laminated gabbros and in the in the adjacent gabbronorites. These appear to have formed by constant-volume or metasomatic replacement of their host rocks. Rocks of this type do not occur at all in the MBG.

6). **Backveined, beerbachitic intrusive sheets.** These, perhaps the most widespread of the minor rock types in the Inner Series, are homogenous, isotropic olivine microgabbronorites and dolerites with porphyroblastic textures. They occur in irregular inclined sheets with irregular margins, and are usually backveined to a greater or lesser extent by the host gabbronorites. The plagioclase phenocrysts present in most are smaller, less calcic and less abundant than those in the gabbronorites. Analogous rocks to these do not occur at all in the MBG.

7). **Ultraporphyritic Dolerites.** These are coarse grained dolerites with an extremely high content of large plagioclase porphyrocrysts but, in strong contrast to the gabbro-norites, contain no orthopyroxene. They also postdate the ultra-high-grade hydrothermal metamorphism characteristic of the Inner Series ( see below ) as do a variety of still later minor intrusions which can plausibly be related to later intrusions defined by Richey et al. ( 1930 ).

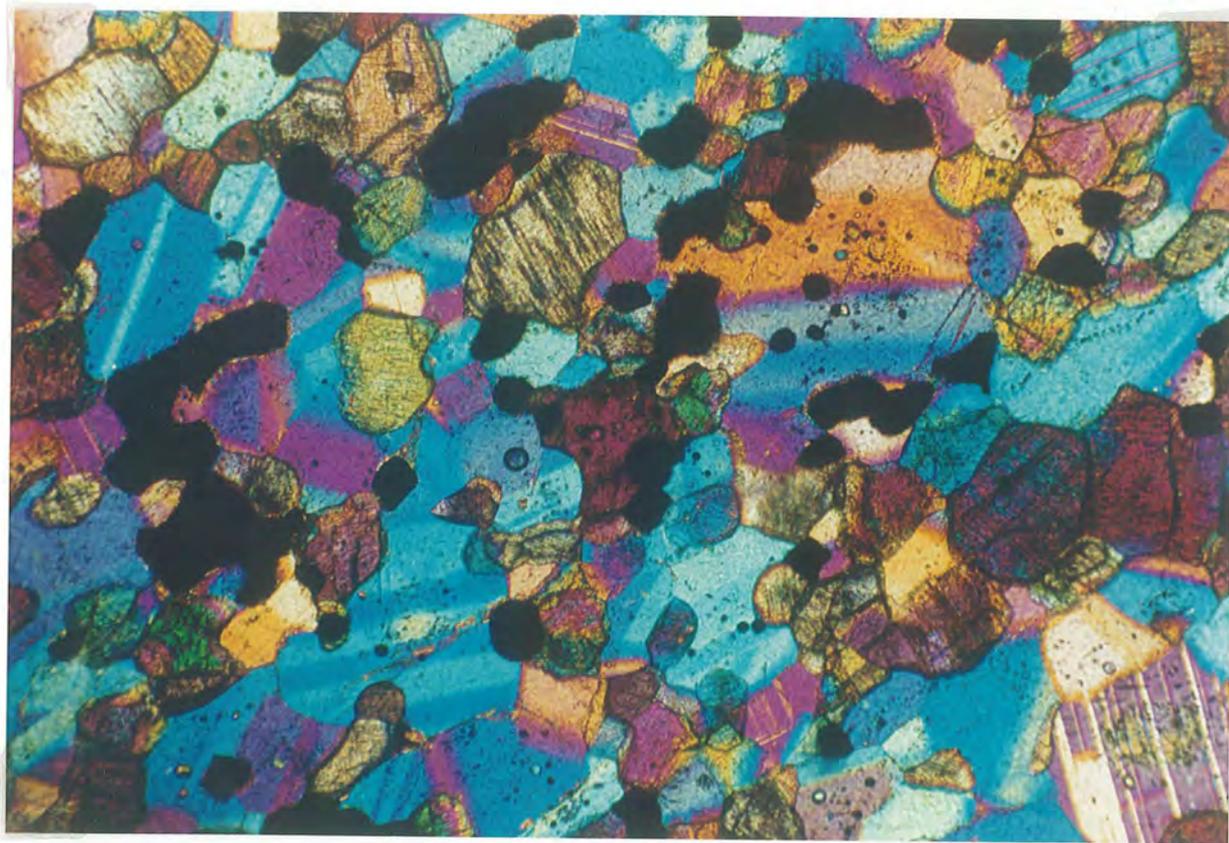
A final feature characteristic of the bulk of the Inner Series is that the rocks have been affected by a phase of very high-grade metamorphism, producing anhydrous mineral assemblages. This is associated with the formation of a tensile fracture network plugged by anhydrous minerals and with the oxidation of olivine. The reactions involved do not occur in the MBG. However the grade of this metamorphism declines towards intrusive contacts between the Inner Series and the MBG and it cannot be used to distinguish the two where it would be useful as a means of doing so.

The Marginal Border Group and the Inner Series are both, therefore, complex intrusive bodies which both contain a variety of rock types, few of which are common to both. As will be seen, they outcrop in well - defined areas, except where sheets of porphyritic rocks cut aphyric rocks of the MBG. These porphyritic rocks may be members of the MBG suite or gabbro-norites belonging to the Inner Series, and in some cases it is necessary to use detailed mineralogical and geochemical analysis to confirm the affinities of such rocks ( see Chapter 6 ).

### **3.3.2. Contact relationships between Marginal Border Group and Inner Series rocks.**

Although the wide variety of differences noted in section 3.3.1 make it easy to distinguish rocks of the MBG from those of the IS in the field and in thin section, few exposures of contacts between the two were found in the field. These are completely lacking on the northern side of the pluton, and very few were found on the southern margin to the east of Beinn nan Codhan. This is because the inferred position of the contact usually lies in a gully of one sort or another, for reasons dealt with below. However, exposures of MBG/IS contacts do occur south and east of Ardnamurchan Point, and above-all around An Acairseid ( see Maps 1 & 4, and Fig. 3.8 ). The three-dimensional structure of the contact can be deduced in this latter area from the distribution of MBG and IS rocks on the western side of Beinn nan Codhan. Two types of contact were observed or inferred: steep sheared or faulted contacts, and flat-lying to shallowly inward-dipping intrusive contacts. The

**Plate 3.71.** Pyroxene - hornfels facies ( although rather retrogressed ) annealed granular shear zone rock from the Ardnamurchan Point fault, just east of the fault itself ( sample 332 ). Primary assemblage is plagioclase + augite + olivine + opaques. 1 -  $\lambda$  plate inserted to show strong preferred crystallographic orientation of plagioclase crystals. Crossed polarisers, width of field of view 3mm.



mechanisms by which the faulted contacts seem to have developed differ, however, between the area from Ardnamurchan Point to Druim na Cloise and that around An Acairseid, so the two areas will be dealt with separately.

The faulted contacts between Ardnamurchan Point and Druim na Cloise which were observed in the field occur along two of a series of NNW - SSE striking fault zones which cut the Hypersthene Gabbro but predate the later large intrusions in western Ardnamurchan ( see section 3.4.7, particularly Fig. 3.29 ). The two faults are the Port Min fault, running between Port Min and Port Choinnich, and the Ardnamurchan Point fault, exposed between the Point and An Acairseid, although it is offset just south of the Point by an east-west fault running through the inlet of Briaghlan. These faults lie on a regional structural trend and seem to be part of regional rather than local, emplacement-related deformation. However, the timing of the bulk of the deformation on them suggests that their orientation may be the result of regional structures being exploited to relieve localised stresses, or the result of the interaction of regional and emplacement-related stresses (see Chapter 7).

Both faults, when exposed, are characterised by the presence of broad zones, 20 – 100m wide, of more or less intensely deformed rocks, cut by large numbers of anastomosing annealed shear zones with granoblastic to poikiloblastic textures. These contain pyroxene hornfels facies mineral assemblages, usually plagioclase + augite + opaques, with small amounts of olivine and/or orthopyroxene, but with some later retrogression to hornblende-hornfelses ( Plate 3.71 ). There is a pronounced preferred lattice orientation of the feldspars in these rocks, with albite-law twin planes lying in the plane of the shear zone. This indicates syn-deformational recrystallisation and plastic deformation at high temperatures, consistent with the post deformational growth of poikiloblastic augite and opaques. The fault zones were subsequently cut by sub-parallel veins containing chlorite and green hornblende and by a number of discrete faults which are invariably concealed in gully floors. The presence of the chlorite- hornblende veins suggests that the reason for this is differential erosion of hydrothermally altered and weakened rocks around these later faults, which seem to have acted as fluid flow pathways. This would also explain the unusually intense alteration of the adjacent pyroxene hornfels facies shear zone rocks. The style of deformation along these faults suggests that they were initially active during a period when most of the rocks they cut were at very high temperatures: it should be noted that where MBG and IS rocks are juxtaposed, the western side of the shear zone is formed by a discrete fault which seems to have juxtaposed rocks deformed at high temperatures against MBG rocks

that do not show evidence of high temperature shearing. By comparison with the faults at An Acairseid described below, this suggests that the faulting was initiated during the emplacement of the Inner Series and formed part of it, but also that movements on the faults continued throughout the subsequent cooling history.

As noted in 3.2.4, the Port Min fault downthrows country rocks, and rocks belonging to the MBG, to the east. Intense parallel jointing on the eastern side of Gharblach Mhor and around 419653 on Druim na Cloise dips east at  $50 - 60^\circ$  and may indicate that the fault has a similar orientation. In contrast, the Ardnamurchan Point fault zone is steeply dipping to subvertical to the east of the Point, dips  $80^\circ$  ENE around 42456514 ( from the preferred orientation of feldspars and the orientation of minor shear zones ) and is very steeply inclined on the northwestern side of An Acairseid. It faults gabbro against gabbro along its entire length but geometric relationships between it and the faults at An Acairseid described below suggest that it is a very steep reverse fault, with the IS rocks to the east being displaced upwards along it.

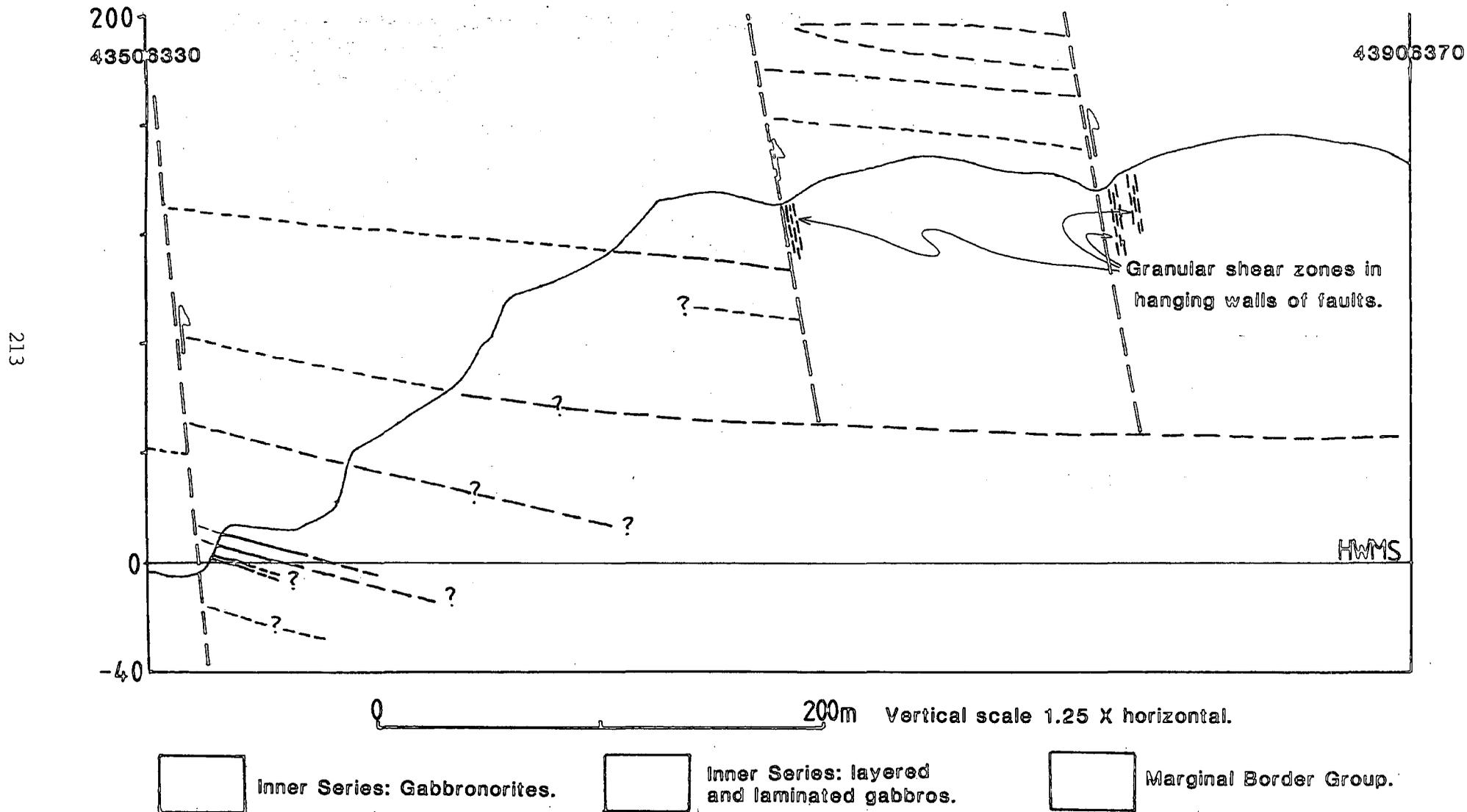
Shallowly dipping intrusive contacts between the MBG and the IS occur along the western coast of Ardnamurchan at Ardnamurchan Point and on the southern half of Gharblach Mhor. To the east of Ardnamurchan Point ( Fig. 3.8 ) Inner Series gabbro-norites occur above granular-textured hornfelsed aphyric microgabbros intruded by porphyritic dolerites in a small area bounded by fault zones. The overlying main contact is not exposed but the overall impression is of one large gabbro-norite body and several satellitic sheets intruding MBG rocks. A similar relationship is developed on Gharblach Mhor, in exposures descending towards sea level as one proceeds south along the ridge. South of about 420649, increasing numbers of porphyritic dolerite and microgabbro sheets appear in hornfelsed granular aphyric microgabbros of MBG type. As one proceeds south, the latter disappear altogether and the entire outcrop is composed of massive, pale-weathering feldspar-porphyritic olivine gabbro-norite south of 42186470. Similar rocks are exposed in the inaccessible sea cliffs to the west of the ridge, suggesting that a north-north-eastward dipping contact between the later gabbro-norite and the overlying hornfelsed MBG rocks underlies much of the area to the west of the Port Min fault. As at Ardnamurchan Point, the gabbro-norite appears to be associated with a swarm of satellite porphyritic dolerite sheet intrusions ( the same relationship also occurs in poorly exposed ground north of Druim na Cloise ). The geochemical affinities of some of these dolerites are problematical, however, and many of them may be MBG suite rocks ( see Chapters 4 and 6 ).

The area around An Acairseid shown in Map 4 contains three distinct groups of faults. One of these is a group of vertical faults, running northeast - southwest ( radial to the Hypersthene Gabbro ), which transfers displacement on NNW - SSE faults to the northwest, including the Ardnamurchan Point fault, to a third group of faults south and east of the inlet which are concentric to the pluton. The radial faults are marked in some cases by granular shear zones, a number of unchilled microgabbroic and doleritic dykes, and above all by intense hydrothermal vein networks running parallel to the faults. The veins commonly contain very low - grade assemblages ( chlorite, serpentine, saponite and iddingsite ) implying that they were open until a very late stage in the development of the faults. Erosion along the latter has produced an impressive valley which obscures the faults inland, but they are well exposed along the coast.

The sense of movement on all of these faults can be constrained with varying degrees of certainty by exposures of rocks offset by the concentric set of faults on the western side of Beinn nan Codhan, immediately east of An Acairseid ( see Map 4 ). A subhorizontal intrusive contact, with only a slight northeastward dip, between older MBG rocks and younger IS rocks, is downfaulted to the south by all four of a series of vertical to steeply NNE-dipping faults which occur to the south of the main fault which separates MBG and IS rocks further east ( Fig. 3.22 ). It should be noted that the geology above the surface along the line of section in Fig. 3.22 is constrained by projection along strike from outcrops on higher ground to the east. The sense of movement on the only exposed fault, at 43566323, is also shown to be reverse by the presence of an en echelon set of chlorite-epidote-rich veins which dip north at a shallower angle than the fault itself, implying upward movement of the northern, hanging wall of the fault. The main, northernmost fault in Fig. 3.22 is also most plausibly interpreted as a reverse fault in the context of the model proposed below for the causes of movement on the others, but direct evidence is lacking. The intrusive contact itself is poorly exposed but is between homogenous MBG microgabbros and a variety of Inner Series gabbroic rocks, which are mainly coarse porphyritic dolerite with large ophitic pyroxenes, and some unusually pale-weathering gabbroic rocks only exposed in cliffs. The presence of the dolerites suggests increased rates of cooling close to the outer contact of the Inner Series. The contact is made more complex in one or two places by the presence of sheets of gabbroic rock intruded into the MBG above the main contact.

The age of movement on these faults is constrained in three ways:

FIG. 3.22. CROSS - SECTION THROUGH THE WESTERN SIDE OF BEINN NAN CODHAN, SHOWING CROSS - CUTTING RELATIONSHIPS BETWEEN CONCENTRIC REVERSE FAULTS AND INTRUSIVE CONTACTS



1). They cut most of the Inner Series rocks in the area but are themselves cut by internal contacts within Inner Series gabbro-norites, most notably in the case of the two northernmost faults. These are not present in the cliffs at the western foot of Beinn nan Codhan and must be truncated by a shallowly dipping intrusive contact within the gabbro-norites at the top of these cliffs between 43556345 and 43666370. This contact is not accessible in this area but if present would be only one of a number of indistinct unchilled planar internal contacts visible in the gabbro-norites between 435635 and 435633. These contacts are themselves cut by the next fault to the south, implying that activity ceased on the northern faults before it did on the latter. Similarly, the southernmost faults are cut by a thin gabbroic sheet in the cliffs around 43586320. However, this gabbro is a pale-weathering, rather fine-grained rock and is atypical of the gabbro-norites, so its significance as a time marker is in some doubt. The presence of these contrasting cross-cutting relationships between faults and various gabbro-norites is also excellent evidence for the interpretation of the internal contacts within the gabbro-norite sequence as intrusive contacts ( see section 3.4.1 ).

2). The northern ( upthrown ) sides of the two northern faults are marked by sub-vertical sheets of granoblastic, sheared and annealed fine-grained basic hornfels like those associated with the faults to the northwest. These contain granular plagioclase + augite + orthopyroxene + opaques + olivine and rare poikiloblastic to interstitial post-deformational augite, implying deformation of the gabbro-norites while they were still at near-solidus temperatures. The downthrown sides of these faults and the three southern faults lack this evidence for high temperature deformation. This implies movement on the faults throughout much of the cooling of the Inner Series gabbro-norites in this area.

3). The faults are older than all of the post - Inner Series minor intrusions that intersect them.

These constraints, and especially the first, imply that these faults were active during the emplacement of the Inner Series gabbro-norites in the An Acairseid area as a series of confluent intrusive sheets ( see Fig. 3.22 and section 3.4.1 ). Uplift of the overlying earlier gabbro-norites and of the MBG / IS contact, as is implied by the geometry of the latter on Beinn nan Codhan, would be a plausible mechanism for creating the necessary space for these intrusions.

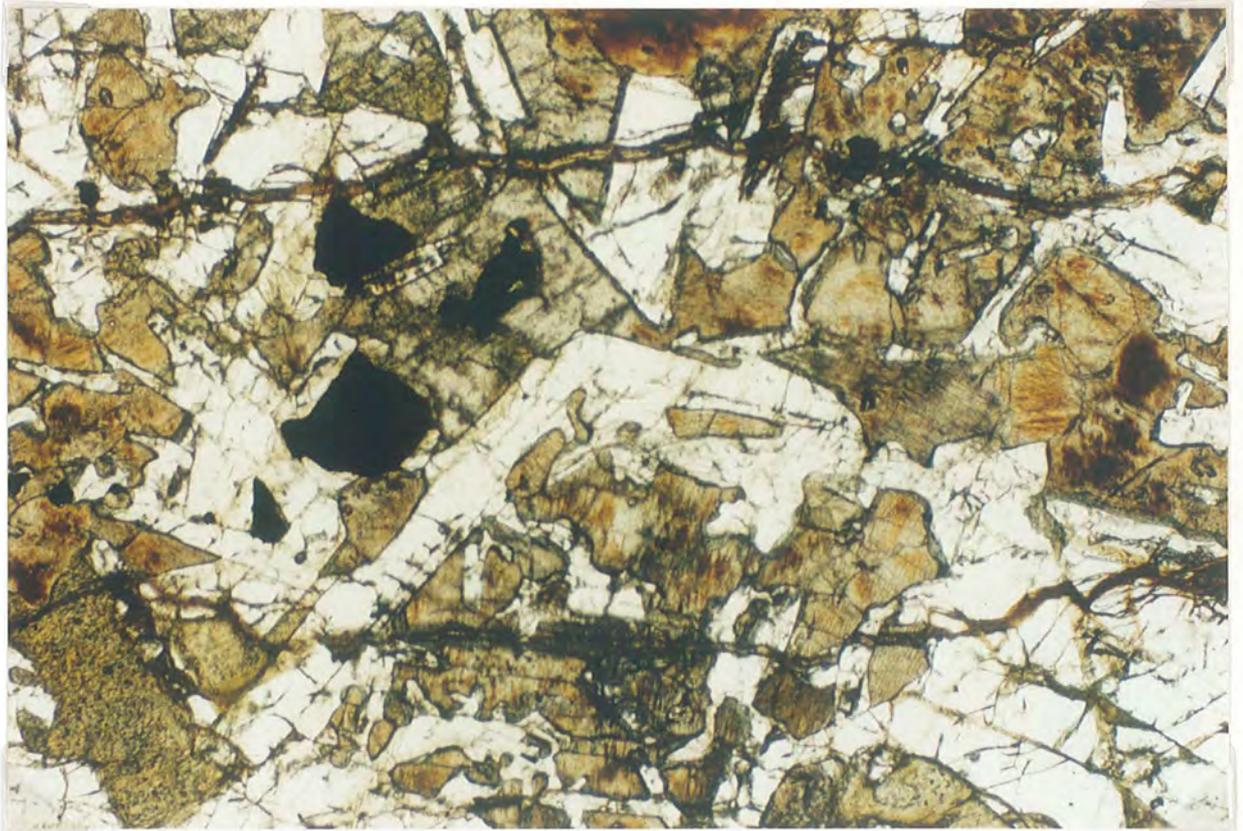
The nature of the radial faults at An Acairseid is best seen in a group of faults on the shore around

433635. These form very prominent fault zones at outcrop ( Plate 3.72A ), cut by numerous sub - parallel vein sets. In thin section they are characterised by the presence of very low grade vein fill assemblages and much low temperature hydrous alteration but retain primary igneous textures and show little sign of penetrative deformation ( Plate 3.72B ), in strong contrast to the concentric faults to the north - west and south - east to which they are connected. They juxtapose MBG and IS rocks, but are terminated only a few tens of metres to the north by a concealed NNW - SSE trending fault which takes up the movement on them. Other radial faults extend further inland but also terminate at NNW - SSE faults. With the exception of the westernmost radial fault, which juxtaposes Marginal Border Group microgabbros and country rock hornfelses at its seaward limit and must therefore downthrow to the northwest, these radial faults juxtapose different gabbroic rocks and their sense of offset cannot be determined directly in the absence of knowledge of the original intrusion geometries. However, the southward downthrowing reverse faults on the eastern side of An Acairseid must terminate at these faults under the bay, which implies that they must downthrow to the northwest. This in turn implies that the large NNW - SSE trending faults, such as the Ardnamurchan Point fault, which take up the movement on the radial faults, must downthrow to the west, away from the interior of the Inner Series. It should be noted that the same constraint does not apply to the Port Min fault, which is not demonstrably connected to the faults around An Acairseid and is of course eastward - downthrowing ( section 3.2.4.1 ). Thus in all cases where a sense of movement can be observed or inferred for faults separating Marginal Border Group and Inner Series rocks, except for the Port Min fault, which cannot be shown to be linked to any other fault in the area, it indicates central uplift of the interior of the pluton. It should be noted that the period of central uplift along faults which this implies is distinctly later than and separate from, that apparent in the shape of reverse faulting in the Maol Bhuidhe area described in section 3.2.5.4.

The northernmost of the concentric faults on the western side of Beinn nan Codhan is probably also a reverse fault, since it is known to have been active only during the emplacement of the Inner Series and is linked to a series of reverse faults to the south ( see above ). It can be traced at least as far east as 44276341, where it outcrops as steeply north-dipping zones of granulitic - textured sheared rocks separating MBG and IS rocks to north and south. The contact between the two groups of rocks is not exposed at all further east, but the outcrops on either side of its inferred position are consistent with a steeply dipping faulted contact.

**Plate 3.72A.** South - west to north - east orientated radial fault zones separating MBG microgabbros ( to right ) from IS gabbro-norites at An Acairseid ( Field location 293; grid reference 43336357 ). The pronounced vertical fabric in these particular faults is defined by a very large number of sub - parallel, anastomosing tensile fractures filled with low - grade hydrothermal mineral assemblages.

**Plate 3.72B.** Microgabbro from within radial fault zone at right - hand side of Plate 3.72A, showing lack of penetrative deformation and clay mineral / serpentine - filled tensile fractures. Sample 293C. Plane - polarised light, width of field of view 3.5mm.



As noted at the beginning of this subsection the contact between the MBG and the Inner Series was not found at outcrop anywhere along the northern margin of the Inner Series east of Sanna Bay. It is marked in areas of generally good exposure by a prominent gully which cuts across the topography and is therefore believed to mark a steep contact. The Inner Series rocks adjacent to it do not show any of the signs indicative of proximity to an external intrusive contact between MBG and IS rocks such as fine grain size and a lack of ultra-high-grade hydrothermal metamorphism ( see section 3.4.6 ) and it is therefore believed to be a faulted contact.

The results of this section can be summarised as follows. Firstly, in contradiction to the conclusions of earlier workers, to the effect that gradational transitions were present between all the components of the Hypersthene Gabbro, a major internal contact has been identified within the pluton, between the Marginal Border Group and the gabbronorites and associated basic rocks of the Inner Series. Secondly, this contact is partly intrusive and partly structural in character. Flat-lying to shallowly inward-dipping intrusive contacts are particularly common along the western and southwestern sides of the pluton, where they are associated with a number of smaller satellitic dolerite sheets intruded into the MBG rocks. The structural contacts are mostly steeply inclined faults, some of which form sets concentric and radial to the pluton, whilst others are part of a swarm of NNW - SSE trending faults which cut western Ardnamurchan. Most of these, with varying degrees of confidence, can be said to be associated with central uplift of the complex. Apart from the sub - vertical radial faults, most of these faults have a reverse geometry. Movement on them took place under conditions varying from near-solidus, pyroxene hornfels facies conditions down to epidote-actinolite hornfels facies conditions at later stages in their activity. At least some of this activity took place at the same time as emplacement of gabbronorite sheets within the Inner Series and may be related to their emplacement.

### 3.4 The Inner Series.

As described in section 3.3, the Inner Series is composed of a large volume of mainly feldspar-phyric olivine gabbronorites ( section 3.4.1 ) and a variety of other rocks which occur in much smaller volumes. These can be divided on the basis of characteristic ages relative to the gabbronorites into a pre - gabbronorite group of mainly layered and/or laminated rocks of varied compositions ( 3.4.2 ), syn- to post- gabbronorite pyroxenites ( 3.4.3 ), and a variety of early and later post- gabbronorite minor intrusions ( 3.4.4 and 3.4.5 ). Distinct metamorphic phenomena in these rocks will be dealt with in section 3.4.6 but as will be seen the distinction between magmatic and metamorphic processes in these rocks is not always clear.

#### 3.4.1 Gabbronorites: cumulate sequence or intrusive sheet complex ?

The distinguishing features of this group of rocks have already been outlined in section 3.3.1. In the past, these rocks have been interpreted as a poorly layered cumulate sequence ( Wells 1954; Skelhorn & Elwell 1971 ) or as a sequence of coarse - grained confluent sheet intrusions ( Wells 1954, 1978 ). Petrographic and field evidence for and against these interpretations falls into the categories of igneous textures, principally the presence or absence of cumulate textures; the presence or absence of layering, lamination and other mesoscopic structures; and the presence of intrusive contacts between gabbronorites and between gabbronorites and certain older rocks. As will be seen the first two types of evidence can be ambiguous because of the uncertainty regarding the mechanisms of formation of textures and structures classically regarded as indicative of crystal accumulation on the floors of large magma chambers. A particularly important issue is whether these structures are cumulate or post - cumulate in origin.

Petrographically, the gabbronorites are mostly coarse - grained but do not contain particularly distinctive or well - developed cumulate textures: in the classical terminology ( Wager et al. 1960 ) they would be described as orthocumulate to mesocumulate. Ophitic to sub - ophitic interstitial pyroxenes are ubiquitous and are up to several centimetres across. The pyroxenes present are mainly augite and bronzitic orthopyroxene; late - crystallised inverted pigeonite occurs in the interstices of a few rocks. The pyroxenes are in general much less deformed than the feldspar crystals which they partially or wholly enclose but lattice straining and subgrain development do occur ( Plate 3.73

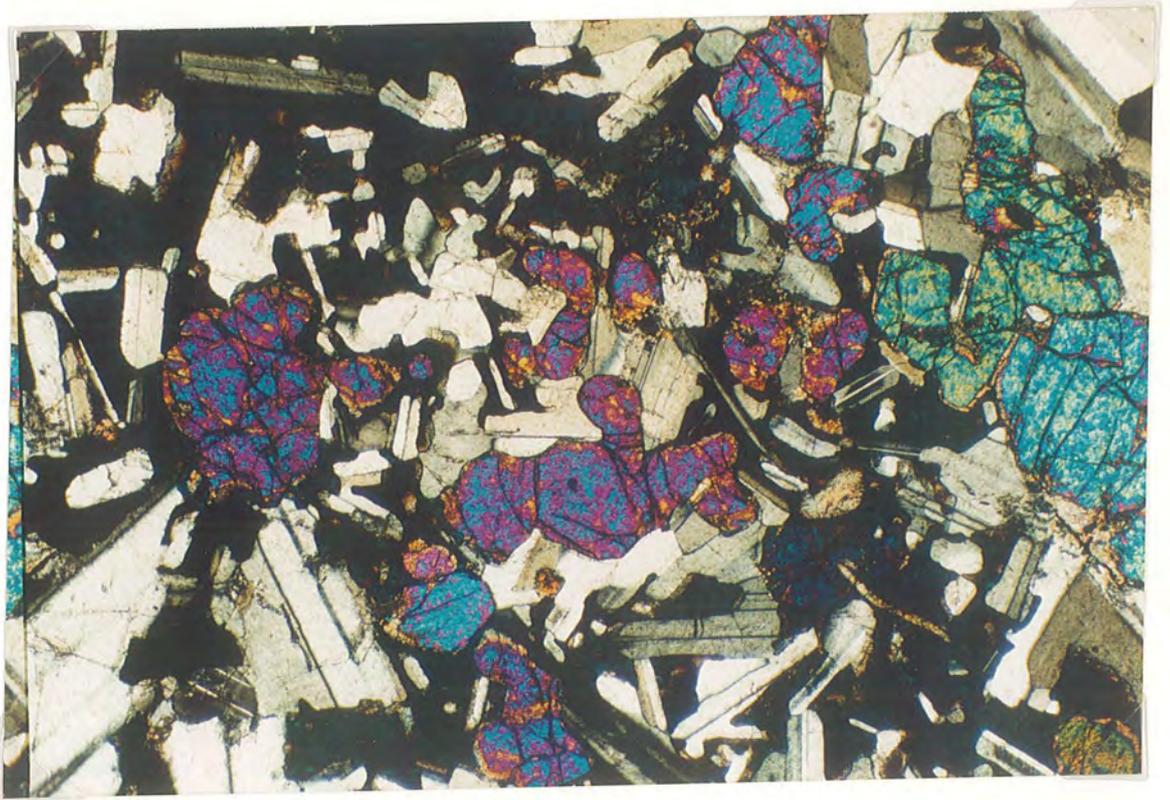
). Olivine, although commonly enclosed ( and possibly corroded ) by orthopyroxene or in a few examples by augite, forms sub - poikilitic to interstitial grains up to 2cm across or clusters of equant grains with more or less similar optic orientations. All gradations between the two occur, in the form of poikilocrysts with subgrains and subdomains developed ( Plate 3.74 ), suggesting that the polycrystalline clusters were produced by breakup of primary poikilocrysts during post - cumulus deformation. A substantial proportion ( up to 25% ? ) of olivine grains in some sections also preserve internal straining and lattice kinking ( Plate 3.75 ). This feature is not normally present in the slowly cooled rocks of normal cumulate sequences where deformation is by the relatively slow process of solution - reprecipitation or textural equilibration ( Hunter 1987 ). Since this deformation does not affect the pyroxenes as strongly, and is therefore syn - magmatic, it seems likely that it reflects higher strain rates rather than lower temperatures during deformation, although the differences in textures of different phases could also reflect different deformation mechanisms. Smaller olivine grains in particular do, however, show rounding against both pyroxenes and feldspar caused by textural equilibration after initial poikilocryst formation. This indicates that solution - reprecipitation processes were important in the deformation of olivine at least.

A few of the olivine poikilocrysts in most of the gabbronorites have large equant central areas which could be cumulus crystals if indeed the rocks are cumulates in the classical sense ( as do certain augites in a few olivine - and orthopyroxene - poor varieties ). Recent textural work on peridotitic cumulates suggests, furthermore, that the true form of olivine primocrysts is in fact dendritic or poikilitic in at least some intrusions ( R.H.Hunter pers comm. ). In this case a large number of the deformed olivine poikilocrysts in these rocks could in fact be accumulative and could only be identified as such from bulk rock compositional evidence ( section 6.2 ). From the textural evidence, however, the main primocryst phase in the gabbronorites is plagioclase.

Plagioclase in these rocks is of two distinct types: porphyrocrysts with multiply - zoned calcic cores, and generally smaller and more elongate groundmass plagioclase. The former make up to 25% of most of the gabbronorites, and substantially more in a few anomalous varieties ( see below ). They are typically 5 to 30mm long and are only slightly elongate, with length to width ratios of 1 to 2 ( exceptionally, up to 3 ). The cores and mantles of these crystals are typically euhedral to subhedral and compositionally uniform ( Plate 3.76 ). In contrast, the outer rims are often irregular, enclose rounded mafic mineral grains similar to groundmass grains of the same minerals

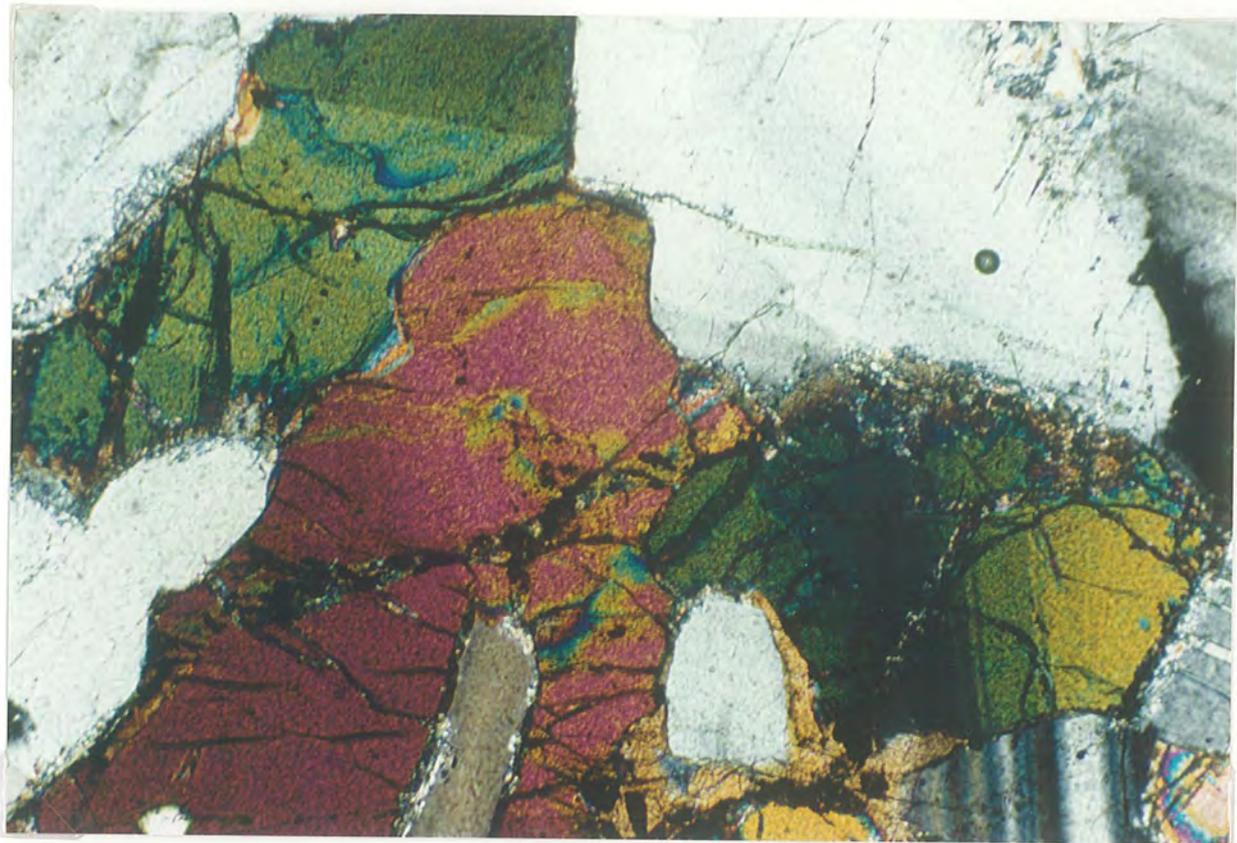
**Plate 3.73.** Strained and corroded plagioclase laths enclosed by a very weakly deformed, largely post - deformational pyroxene oikocryst. Gabbronorite, sample 204. Crossed polarisers, width of field of view 3mm.

**Plate 3.74.** Olivine poikilocryst, with subgrains present within it, and strained plagioclase crystals enclosed in entirely undeformed augite oikocryst ( at extinction ). Both the olivine and, to a lesser extent, the feldspars, show texturally equilibrated grain boundaries. Fine - grained gabbronorite, sample 296. Crossed polarisers, width of field of view 3mm.



**Plate 3.75.** Granular to sub - poikilitic olivine grains showing internal deformation and kink banding. Sample 287B2 ( gabbronorite ). Crossed polarisers, width of field of view 1.5mm.

**Plate 3.76.** Subhedral plagioclase porphyrocryst with unzoned core, distinct outer core zone, and continuously zoned rim. Note indentation by adjacent plagioclase lath, which cuts across the zonation implying late - stage dissolution. Patchy re - equilibration of the core to feldspar with the same composition as the rim has also occurred. Sample 165 ( gabbronorite ). Crossed polarisers, width of field of view 3mm.



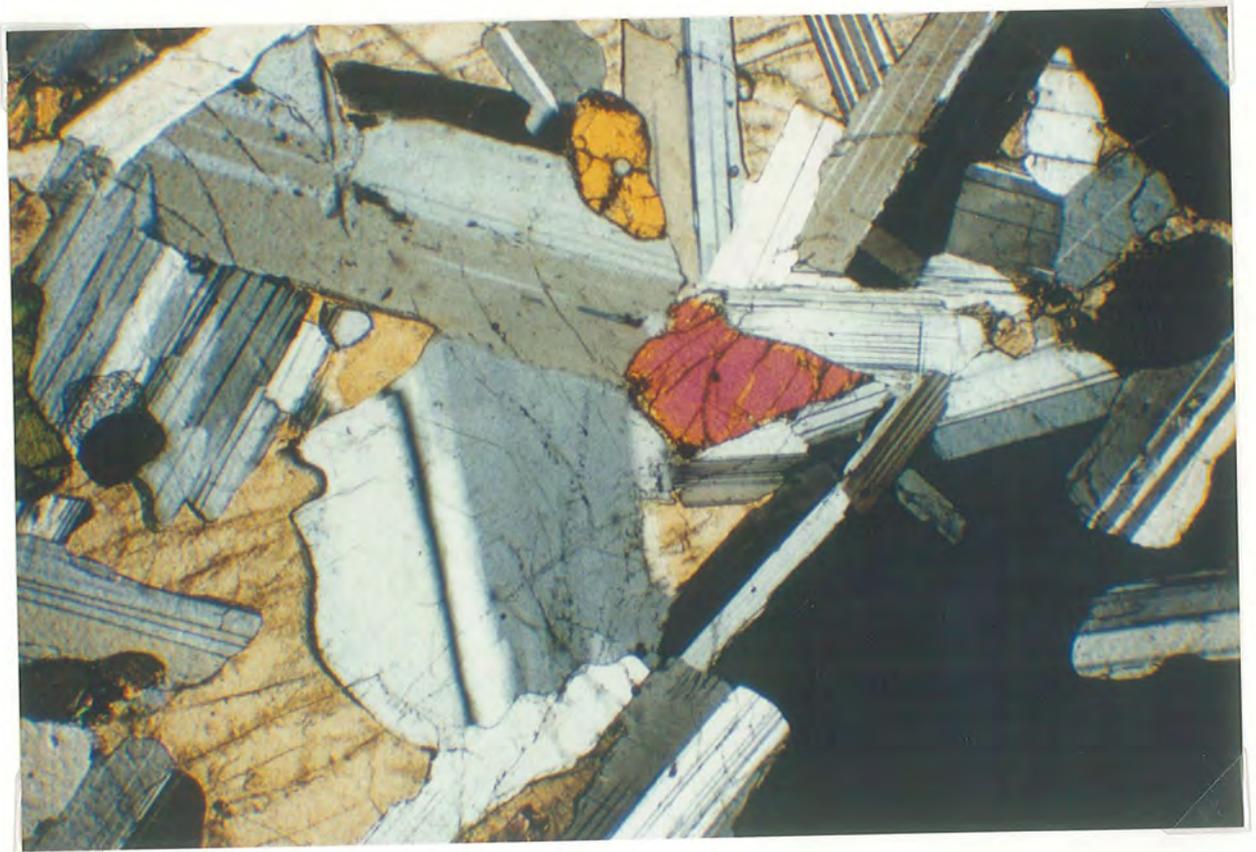
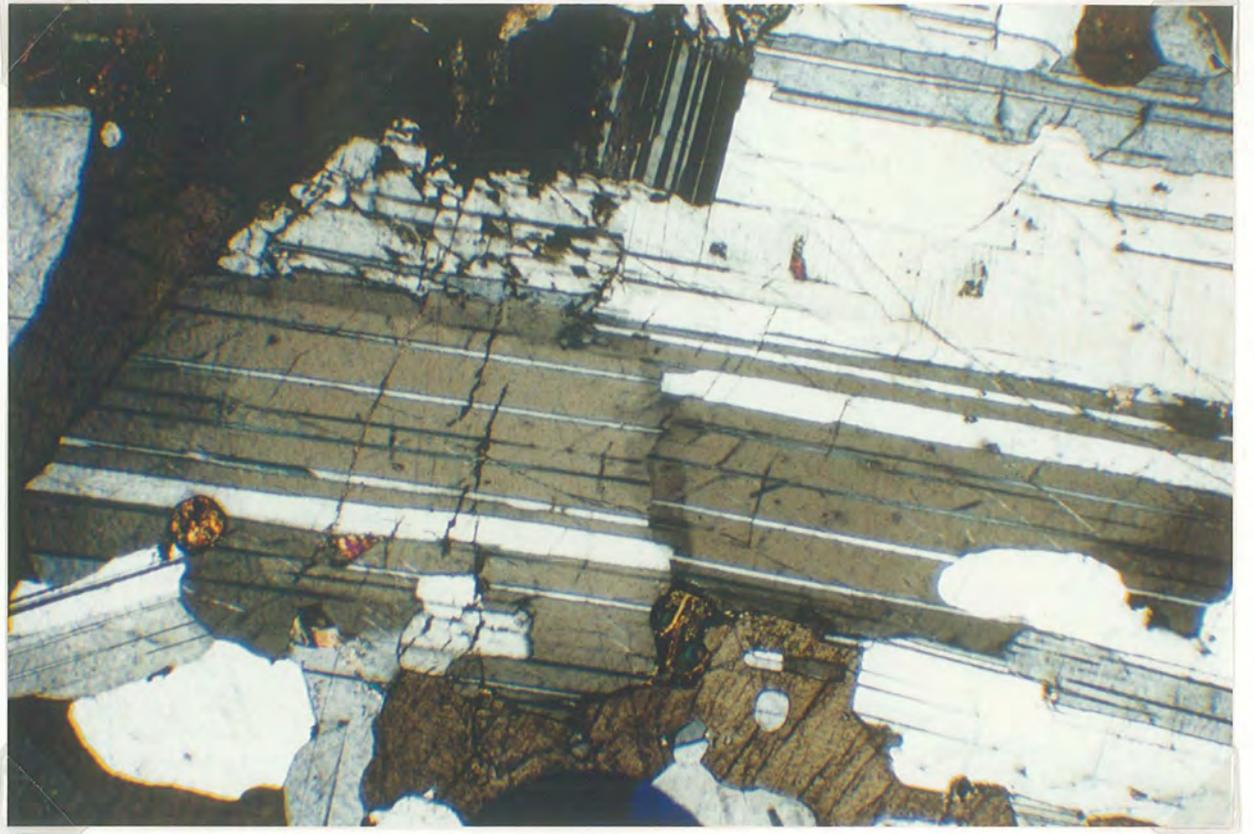
and are strongly zoned to groundmass compositions. This zonation is often truncated by mafic poikilocrysts, suggesting localised dissolution of the feldspars during growth of the latter. In some cases the mafic grains cut across the cores and inner rims of the plagioclase porphyrocrysts, indicating that these were also affected by localised feldspar dissolution processes. Replacement of the feldspar cores, by feldspar with the same lattice orientation but with groundmass feldspar rim composition occurs along fractures, twin planes and subgrain boundaries, and around mafic or opaque mineral inclusions. The composition of the secondary feldspar in this texture ( see Chapter 6 ) suggests that it formed at a late magmatic stage, or possibly just below the solidus. The groundmass feldspars, which vary from subhedral laths showing rounding and resorption against adjacent mafic phases to highly strained, more equant and anhedral grains ( especially abundant in the coarser rocks ) show slight but distinct core - to - rim zonation.

The finer - grained gabbroites typically have relatively well preserved doleritic textures with large and prominent ophitic pyroxenes enclosing slightly rounded but undeformed plagioclase laths. In the much more abundant coarser - grained varieties, the feldspars record variable but often considerable deformation. This takes the form of lattice bending, producing anomalous extinction and bending of twinning, and suturing at grain boundaries and twin planes. Larger feldspar grains, particularly the calcic cores of the porphyrocrysts, contain subgrains and annealed brittle fractures ( Plate 3.77 ). Dissolution at grain boundaries, cutting twin planes and zonation, is also common but textural equilibration, which would be characterised by the development of polygonal plagioclase - plagioclase grain boundaries, is only ever imperfectly developed.

The timing and temperature of the deformation in the feldspars can be constrained by their textural relationships with the mafic phases. The olivines do not normally enclose strained feldspars and are either deformed or have undergone subsequent recrystallisation ( see above ). This suggests that the original oikocrysts were either syn- or pre- deformational. In contrast the pyroxenes show only weakly deformed primary ophitic textures and contain essentially undeformed exsolution lamellae, whilst enclosing intensely strained feldspars. Opaque grains also form undeformed interstitial oikocrysts. The bulk of the deformation is therefore constrained to have taken place at an advanced stage of crystallisation, when the feldspars formed an interlocking crystal network and the olivines had also crystallised , but before crystallisation of most of the pyroxenes and of the opaques. Since these have good poikilitic - interstitial textures it seems most likely that the deformation in the feldspars

**Plate 3.77.** Plagioclase porphyrocryst cut by annealed fracture. Note that re-equilibration of the core of the porphyrocryst to the same composition as that of the rim has occurred along this fracture. Sample 164 ( gabbro-norite ). Crossed polarisers width of field of view 3mm.

**Plate 3.78.** Strained equant plagioclase grain with albite - law twin planes at a high angle to a crude plagioclase lamination which runs horizontally across the plate. The greater development of internal deformation characteristic of grains with this orientation suggests that the lamination formed by a process of late - stage compaction. Unusually, the plagioclase laths in the augite oikocryst at the top of the slide are randomly orientated and largely undeformed, suggesting that this oikocryst grew before the lamination developed. Sample 287B2. Crossed polarisers, width of field of view 3mm.



took place above the solidus. The deformation is pervasive throughout the coarser gabbronorites but is less well developed in finer - grained varieties. This widespread development suggests that the deformation does not reflect localised fault - zone deformation ( see section 3.4.7 for a discussion of faulting in the Inner Series ).

On a mesoscopic scale, most of the coarser gabbronorites show a weak to very weak feldspar lamination defined by the more elongate feldspar laths. This is often more apparent in thin section than at outcrop. The lamination is **not** reflected in a preferred orientation of the feldspar porphyrocrysts and is more a morphological orientation than a preferred lattice orientation. Those grains with albite twin law twin planes ( which are parallel to the long axes of the elongate laths ) at high angles to the lamination tend to be more equant, to have irregular grain margins and to show internal lattice straining ( Plate 3.78 ). These features suggest that the lamination formed by growth of crystals with z - axes in the plane of lamination, and deformation and dissolution of crystals with z - axes at high angles to the lamination. These features imply that the lamination is not a cumulus structure but developed as a result of supra - solidus deformation in a crystal mush. Post cumulus processes of this type have been proposed by Maaloe ( 1976 )( dissolution of grains at high angles to lamination ) and by Mathison ( 1987 )( growth of grains in the plane of lamination ). The development of lamination in this way probably reflects compaction of a crystal mush, either under its own weight in the case of cumulates on the floor of a magma chamber ( McKenzie 1987 ) or under the weight of older rocks in the case of sills intruded into partially molten rocks. It does not necessarily indicate that the gabbronorites are normal cumulate rocks.

Another feature normally found in cumulates which also occurs in the gabbronorites is an almost complete absence of visible evolved interstitial material ( apatite, alkali feldspars, zircon, granophyric intergrowths etc. ): the interstices of the crystal fabric are occupied instead by pyroxenes and opaques. The coarse grain size of the rocks and the vanishingly small amounts of late - crystallising material present make estimation of its abundance by point counting impracticable: the alternative approach, of estimating the abundance of trapped residual melt by geochemical analysis, is followed in Chapter 6. The main occurrences of late - crystallising material in the gabbronorites are the augite - rich gabbroic pegmatoid pods and veins which occur in many of the gabbronorites. As with the development of lamination, the depletion of rocks in late - crystallising components can occur by cumulus processes or by compaction and melt expulsion, interstitial melt migration in temperature

gradients or other crystal mush processes, and is not necessarily an indicator of formation by primary cumulus processes.

There is therefore much petrographic evidence in the gabbronorites for the operation of the **postcumulus** processes of crystal mush compaction and melt expulsion at a time when they were partially molten crystal mushes composed of a deforming network of plagioclase and olivine crystals in continuous contact and an interstitial melt which mainly crystallised pyroxenes, plagioclase and opaques. In contrast, good evidence for **cumulus** processes is absent except for an upward decrease in plagioclase porphyrocryst abundance in a few thin gabbronorite bands, which may reflect accumulation of these crystals by crystal settling. Most outcrops of gabbronorite show uniform distributions of these crystals, at least within the individual sheets which make up many of these outcrops.

Structures and rocks resembling those of true cumulate rocks have however been described from the Inner Series ( Wells 1954, Skelhorn & Elwell 1971 ). Most of these occur in other types of Inner Series rocks ( see sections 3.4.2 and 3.4.3 ) but some occur in gabbronorites, especially in gabbronorites at the southern end of a series of outcrops of laminated gabbros on the coast between Sanna Point and Sanna Bay ( see Fig. 3.25 ) around 44076998, and in a porphyritic gabbronorite unit **within** the laminated rocks around 44057000. In both cases there is evidence to suggest that the gabbronorites form thin sheets with intrusive margins, in the shape of veining of **overlying** rocks and the presence of finer grained margins which may have resulted from increased rates of cooling and hence nucleation densities at the margins of intruded sheets. This carries the implication that the laminated bands and other structures described from these rocks are not in fact cumulate structures and have to be re - interpreted, given the intrusive nature of the gabbronorite sheets which contain them.

The rocks in this area form a continuous series of outcrops, aligned north - south, in which planar or sheet - like structures dip south at  $15 - 30^\circ$  . The structures can be traced down dip for tens of metres, and inland for similar distances, and within these limits they are to a large degree laterally uniform. This allows the sequence to be represented as vertical logs ( Figs. 3.23 and 3.24 ) although many of the contacts are discordant at low angles as well as cross cutting the older rocks on small scales. These discordances are important in that they confirm the intrusive character of the gabbronorites and the secondary origin of the pyroxenites ( see section 3.4.3 ) in these logs. A marked feature of

Fig. 3.23. VERTICAL LOG THROUGH PORPHYRITIC GABBRONORITE SHEETS AT SOUTHERN END OF EXPOSURES OF LAMINATED GABBROS SOUTH OF SANNA POINT. SEE Fig. 3.25 FOR LOCATION.

ENLARGED SECTIONS OVERLEAF ( Fig. 3.24 ).

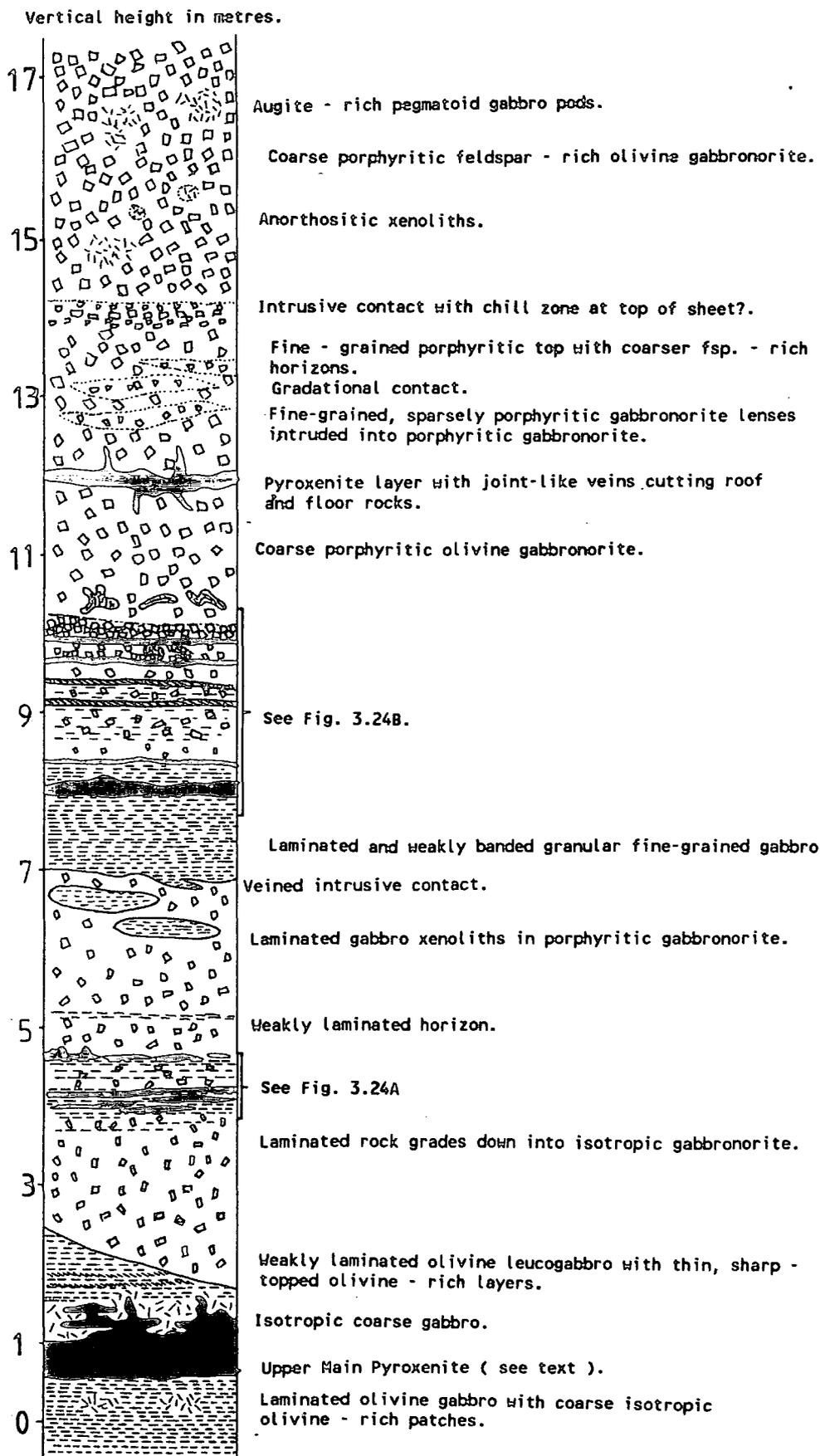
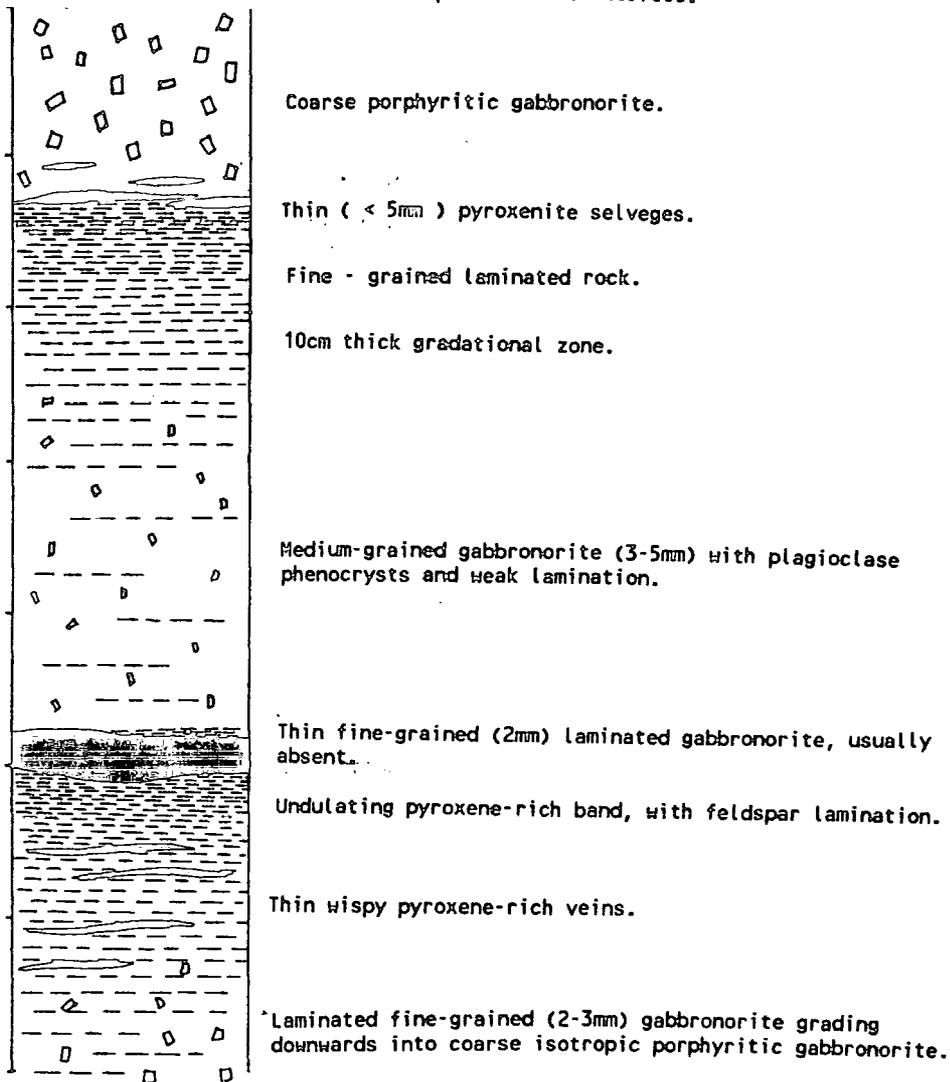
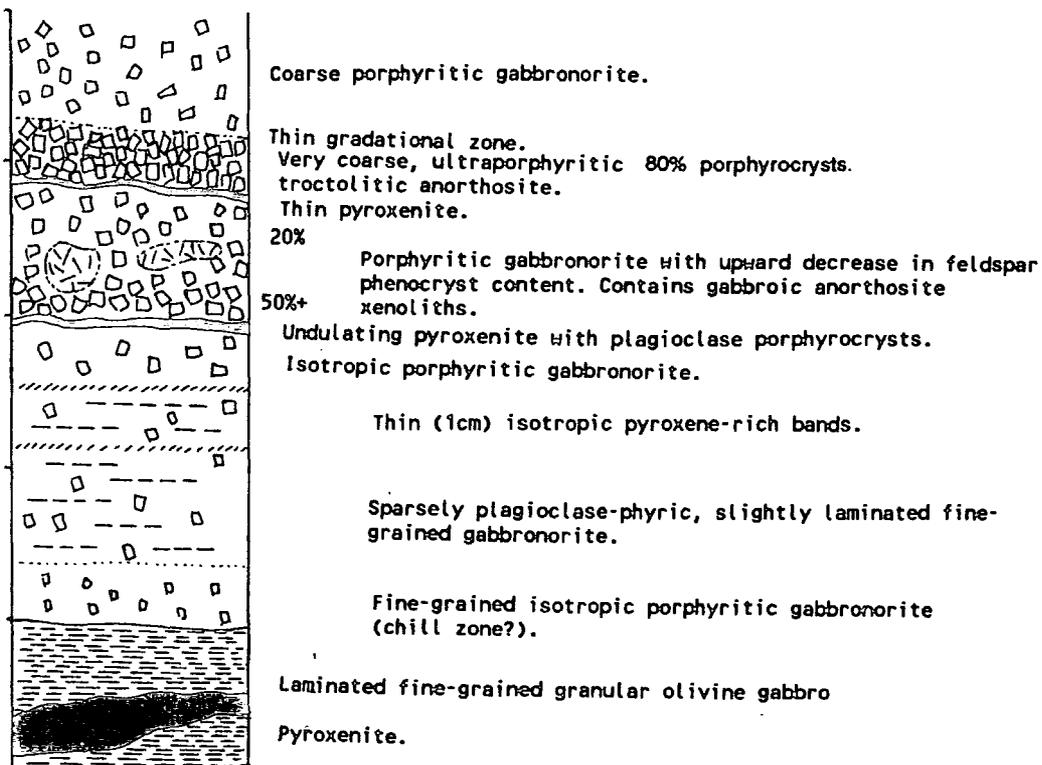


Fig. 3.24. DETAILED LOGS FROM Fig. 3.23.

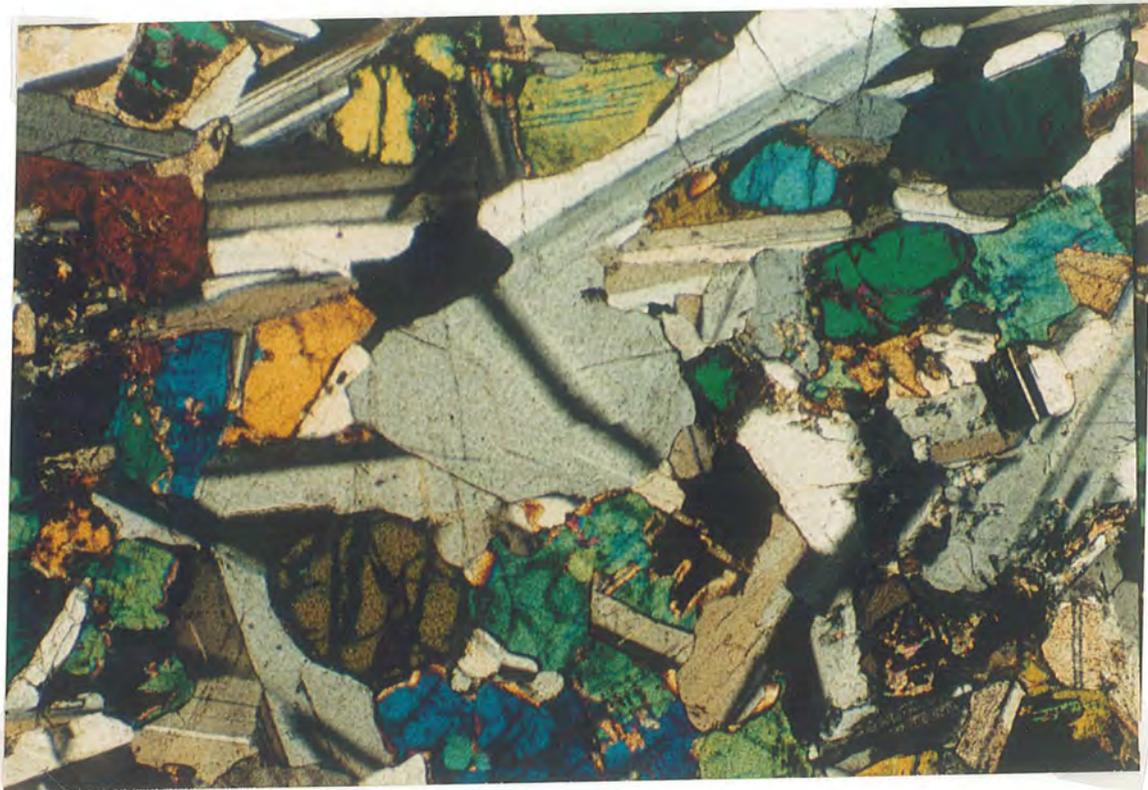
A. 3.8 to 4.5m ( scale 1cm=5cm ): exposed around 44057000.



B. 8.0 to 10.5m ( scale 1cm = 25cm ): exposed around 44066998.



**Plate 3.79.** Highly strained and internally fractured anhedral plagioclase grain with twin planes at a high angle to a well - developed plagioclase lamination. From a well - laminated band within otherwise near - isotropic gabbro, sample 315C. Crossed polarisers, width of field of view 3mm.



Figs. 3.23 and 3.24 is the association of lamination in the gabbronorites with secondary pyroxenites. Elsewhere in the same sequence, particularly in the laminated gabbros, the pyroxenites are associated with the destruction of lamination ( see 3.4.3 ) so the relationship between the two is not a simple one.

In thin section the well - laminated units in the gabbronorites, like the normal weakly laminated gabbronorites ( see Plate 3.78 ) contain short, highly strained grains with their z - axes at high angles to the lamination ( Plate 3.79 ). In addition the lamination is symmetrically wrapped around olivine oikocrysts **which enclose randomly orientated plagioclase laths**. This implies growth of these oikocrysts **prior** to compaction in a direction perpendicular to the lamination which caused it to be bent around the oikocrysts. Although such oikocrysts may not necessarily have formed in the interstices of a crystal mush, the symmetric deflection of the lamination around them strongly implies that the lamination is a secondary or post - cumulus structure, as argued by Mathison ( 1987 ) in the case of similar deflections of lamination around pyroxene oikocrysts. The pyroxenes in the laminated units in the gabbronorites are usually broken up into subgrains, implying that deformation in these laminated rocks was more intense even after pyroxene crystallisation than in the normal gabbronorites, and may therefore have continued to lower temperatures as well ( see Chapter 6 ). Where interstitial - poikilitic augites are preserved, however, they do not deflect the lamination, indicating that they grew after its development.

Modal layers in the gabbronorites in Figs. 3.23 and 3.24 are of two types. The most numerous are thin sub - conformable pyroxene- and olivine- rich isotropic veins and selvages which appear to be incompletely developed secondary pyroxenite layers ( section 3.4.3 ) and layers with unusually high contents of plagioclase porphyrocrysts, as noted above. Other apparent 'layers' in the sequence appear to be the upper and lower margins of sheet intrusions and some late gabbroic segregation veins and fine - grained gabbronorite sheets.

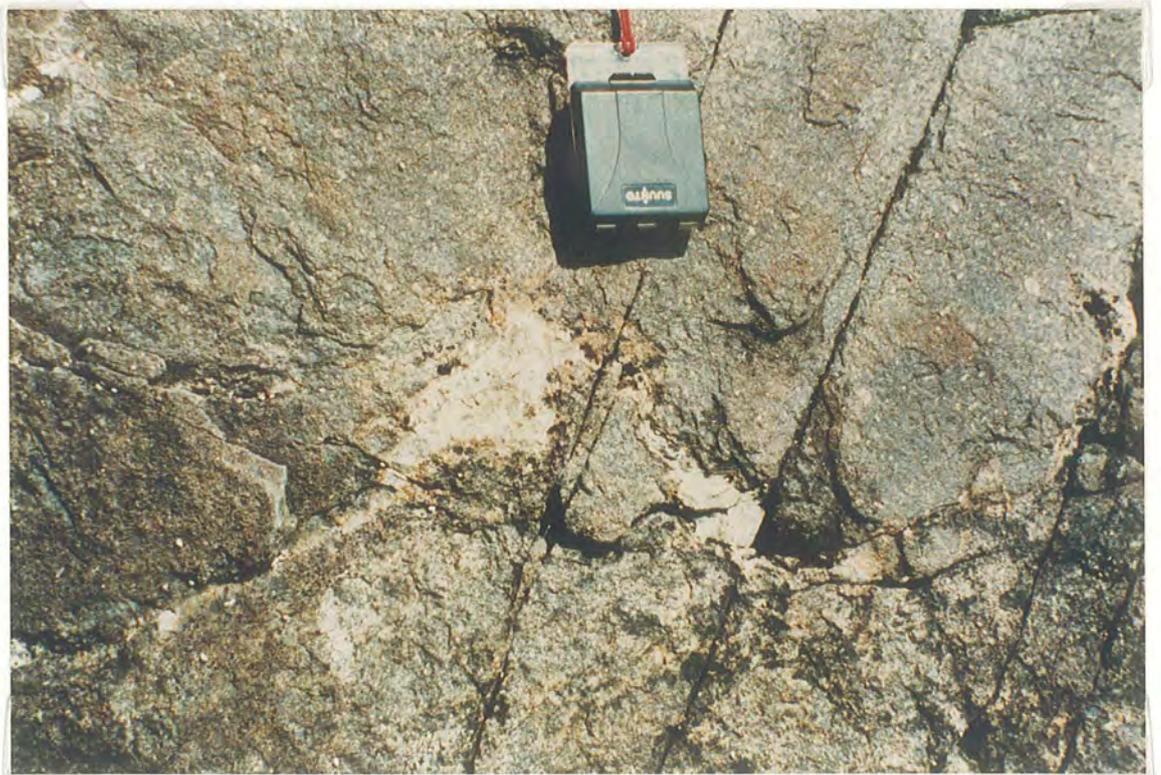
Elsewhere in the Inner Series cumulate - like structures are even less abundant in the gabbronorites: almost all occurrences of layering and lamination are found in older and mineralogically distinct gabbros and granular rocks which are demonstrably cross - cut by the gabbronorites ( see section 3.4.2 ). Laminated units do occasionally occur in the gabbronorites. At one such locality, ( Grid Ref. 44196943 ) the laminated band is truncated by the contact between the host gabbronorite and a

screen of older rocks, suggesting that it is not simply the product of compaction and recrystallisation of the host gabbro-norite, although the original model for igneous lamination ( crystal settling in a laminar flow ) cannot explain it either. The contacts between the gabbro-norites and the laminated and layered rocks are typically planar and shallowly inward - dipping overall, but small - scale irregular gabbro-norite veins ( such as those in Plate 3.80 ) are common. In the example shown in Plate 3.80, the overlying laminated anorthosites are veined and brecciated by the **underlying** gabbro-norites which are therefore the younger rocks. This, the reverse relationship to the upward - younging sequence which must be present in conventional cumulates is commonly developed in the Sanna area ( see Fig. 3.23 for examples ). Its occurrence forms strong evidence in support of the interpretation of the gabbro-norites as a sequence of confluent sheet intrusions. Stopping and spalling of blocks and slabs from the walls of such intrusions, as is seen in Plate 3.80, would be a plausible mechanism for the incorporation into the gabbro-norite magmas of the tabular to rounded xenoliths, of granular to laminated banded and coarse poikilitic to granular isotropic leucogabbroic and anorthositic rocks, that occur widely in the gabbro-norites. These are particularly common in the Sanna area, and also adjacent to outcrops of laminated rocks at An Acairseid ( see 3.4.2.2 ) which are also cut by gabbro-norites. Most of the xenoliths at this locality are anorthositic or near - anorthositic, and occur at the centre of irregular gabbroic vein networks ( Plate 3.81 ). Similar veins occur around many of the isotropic xenoliths at Sanna. This suggests that even the anorthositic xenoliths have undergone partial melting, and that extensive assimilation of less refractory, more nearly cotectic rocks may have taken place. Anorthositic rocks may be over - represented in the xenolith population because they would be preferentially preserved even in magmas more primitive, and hence hotter, than those they crystallised from ( see Chapter 6 ).

Intrusive contacts within the gabbro-norites are not obvious, because of the close similarity of the two rock types involved and the absence of pronounced chilling or hornfelsing, but are present in large numbers. The largest example identified is the one which cross - cuts the faults at An Acairseid, described in section 3.3.2. Indistinct but sharp contacts are visible in individual outcrops, such as that shown in Plate 3.82 ( which is the lower contact of the sheet whose upper contact is that shown in Plate 3.81 ), and in the cliffs at An Acairseid ( Grid Ref. 435635; Map 4 ). These gabbro-norite - gabbro-norite contacts are planar and consistently dip towards the centre of the Hypersthene Gabbro. Dip angles are usually in the range 15° to 25°. Similarly orientated boundaries between subtly different gabbro-norite variants, such as xenolithic and xenolith - free types, and coarser and finer

**Plate 3.80.** Pre - gabbronorite anorthosites ( at top of outcrop ) veined and brecciated by the underlying gabbronorite sheet, implying that the latter is intrusive. Field location 287C, grid reference 44136978 ( in the middle of the eastern group of anorthosite outcrops on the northern side of Sanna Bay, Fig. 3.25 ).

**Plate 3.81.** Anorthositic xenolith in gabbronorite, field location 275/1, An Acair-seid. The xenolith is about 4cm across. Grid reference 43536333 approx.



**Plate 3.82.** Planar contact at base of gabbro-norite sheet, Sanna. Note truncated xenoliths in older underlying gabbro-norite sheet at northern (left-hand) end of outcrop and weakly developed chill zone at base of sheet. Field location 287D, grid reference 44136980 approx.



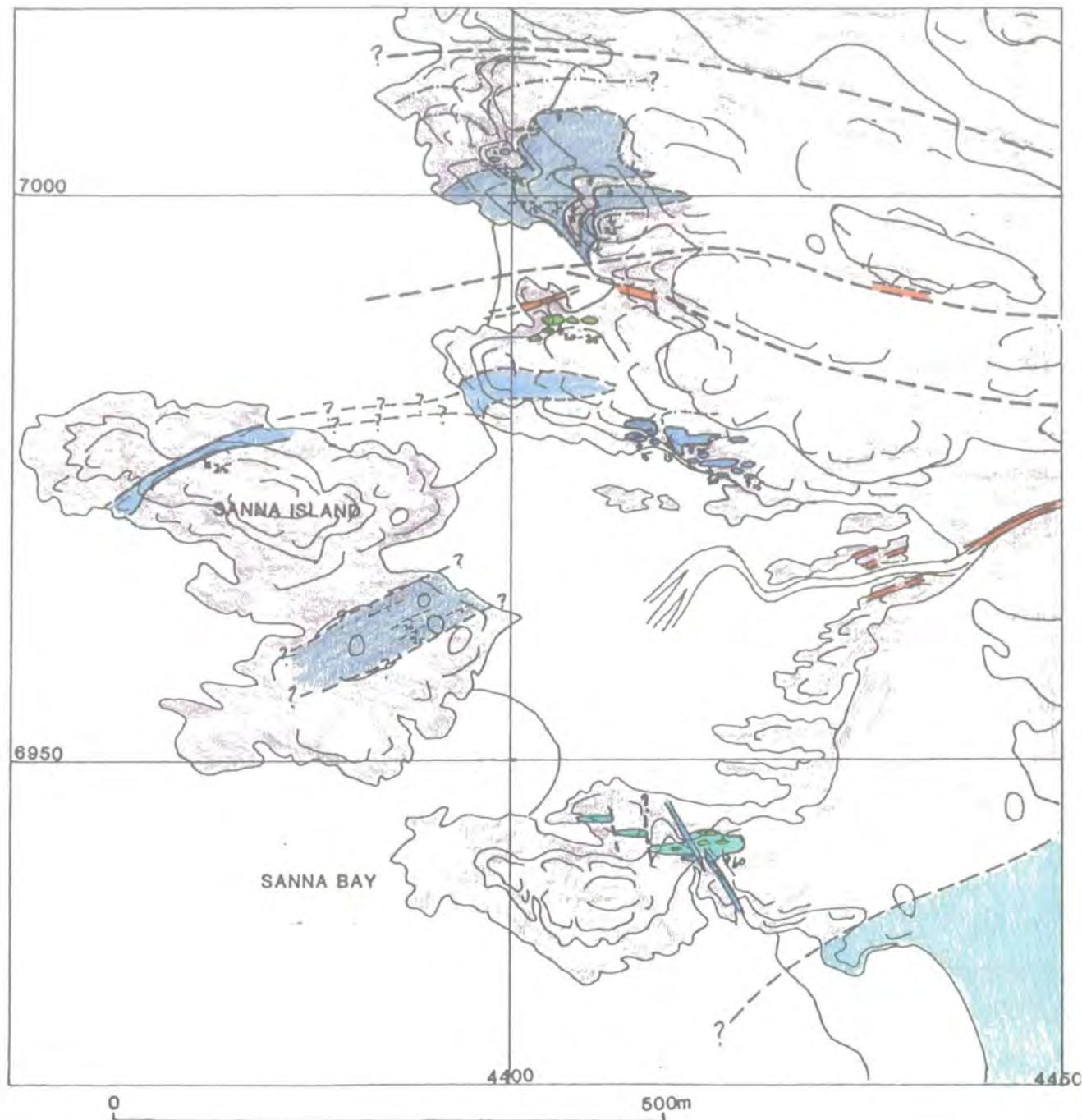
grained rocks, can be mapped over distances of as much as a few hundred metres. A particularly good example of this occurs north of the laminated gabbro screens at Sanna ( Fig. 3.25 ). The top of this sequence includes a 3m thick unit of fine - grained gabbronorite immediately below the laminated gabbros, and about 3m of gabbronorite with a crude banding defined by subconformable pegmatoid gabbroic sheets just below that. This sequence is reminiscent of the upper part of many differentiated sills, in which migrating interstitial residual melts accumulate beneath the roof of the sill, and the sequence of gabbronorites from this outcrop to about 60m to the north may represent a single weakly differentiated sheet intrusion about 10 to 15 metres thick.

It therefore seems that some at least of the gabbronorites form sheet - like intrusions with shallow inward dips, emplaced into rocks already at near - magmatic temperatures and therefore lacking good chilled margins. These sheets occur both at the periphery of the Inner Series, producing the shallowly - dipping intrusive contacts described in section 3.3, and in its interior. As noted at the beginning of this section, however, the bulk of the gabbronorites are monotonous near - isotropic coarse grained rocks with few features characteristic of any particular mode of origin. The lack of primary cumulate structures in them and their close textural and microstructural similarities to those gabbronorites which do occur in distinct intrusions suggests that all the gabbronorites form a sequence of confluent inclined sheet intrusions. They must, however have been subjected to crystal mush compaction, residual melt expulsion and recrystallisation at unusually high strain rates to produce the petrographic features described above, implying that each must have formed part of a much thicker body of partially molten rock, with dimensions of the order of those of the Inner Series as a whole.

#### **3.4.2 Pre - gabbronorite layered and laminated rocks.**

Rocks with well - developed igneous lamination and modal layering occur mainly in three restricted areas of the Hypersthene Gabbro: between Sanna Point and Sanna Bay ( inclusive ); on and around Beinn na Seilg, to the north - west of Kilchoan ( see Map 1 ); and at An Acairseid ( Map 4 ). In addition, isolated well - layered and/ or laminated xenoliths occur throughout the Inner Series. Perhaps the most striking feature of these early rocks, as a group, is their very wide compositional, mineralogical and textural diversity. Individual lithologies are, to varying extents, confined to particular regions of the Inner Series and it is therefore convenient to consider each of the main areas

FIG. 3.25. XENOLITHS AND SCREENS IN THE INNER SERIES, SANNA.



- |   |  |   |   |
|---|--|---|---|
|  | Great Euclite.   |  | Coarse anorthosites and anorthositic gabbros.                             |
|  | Hybrid granophyre/dolerite<br>( part of Sgurr nam Meann intrusion? ).<br>Inner Series. |  | Finely-banded eucrites and granular<br>apatite-eucrites and anorthosites. |
|  | Porphyritic Gabbronorites.   |  | Marginal Border Group.  |
|  | Observed internal contacts within<br>the porphyritic gabbronorites.                    |  | Dip of lamination and/or layering.  |
|  | Laminated gabbros.   |  | Dip of intrusive contacts.  |
|  | Laminated microgabbronorite  |  | Inferred subvertical faults.  |

of occurrence of these rocks in turn.

#### 3.4.2.1 Sanna Point and Sanna Bay.

The coastal area at Sanna, between the outer contact of the Inner Series and its inner contact against the much later Great Eucrite, contains a wider variety of pre - gabbronorite rocks than any other. In part this may be due to the juxtaposition of originally vertically separated Inner Series rocks along concentric faults which cut the area ( Fig. 3.25 ).

The largest and best - exposed group of such rocks occurs close to the outer edge of the Inner Series, in an area 150m by 200m ( east - west ) centred on 44057002. These are a gently south - dipping series of rather granular, fine - grained ( grain size 2 - 3mm ), well - laminated and weakly layered, orthopyroxene - poor to -absent and near - aphyric olivine gabbros. They are intruded by porphyritic gabbronorite sheets, and have the overall form of two sub - parallel, sub - conformable south - dipping screens ( Fig. 3.25 and Fig. 3.23, above ). They are also cut by two thick and numerous smaller secondary pyroxenite sheets which are dealt with separately ( section 3.4.3 ). Later fine - grained beerbachitic dolerite sheets are also present ( 3.4.4 ). Similar slightly granular laminated rocks occur on the south side of Sanna Island in barnacle - and seaweed - covered outcrops which it was not possible to examine in detail.

The plagioclase lamination in the outcrops to the north of the bay dips south, towards the centre of the pluton, at about 15°, slightly less than the dip of the enclosing contacts with the gabbronorites. The feldspars in the plane of lamination are elongate subhedral laths, while the small proportion of crystals with albite - law twin planes at a high angle to the lamination are equant, anhedral and highly strained. A very small proportion of the plagioclase crystals ( no more than 1 - 2% of the whole rock ) are phenocrysts which, like those in the gabbronorites, have unzoned, rounded and corroded cores, some of which have an unzoned mantle in addition to the zoned rim. The lamination in these gabbros, like that which is intermittently developed in the gabbronorites, appears to have developed by deformation of a crystal mush. This is confirmed by the relationship of the lamination to slightly irregular but laterally continuous troctolitic layers which occur at intervals of 5 to 30cm through most of the laminated gabbro sequence. These layers are only a centimetre or so thick and contain poikilitic, slightly strained olivine grains up to 2cm long. These crystals enclose elongate,

randomly orientated plagioclase laths and deflect the lamination in the enclosing rock where they protrude into it ( Plate 3.83 ). The deformation in these rocks also affects the pyroxenes in the rock, which are mostly reduced to irregular subgrains, suggesting that it continued to a more advanced stage of crystallisation. However, where interstitial - ophitic augite grains are preserved in the laminated rocks, they do not deflect the lamination, suggesting that they post - date that part of the deformation which resulted in formation of lamination.

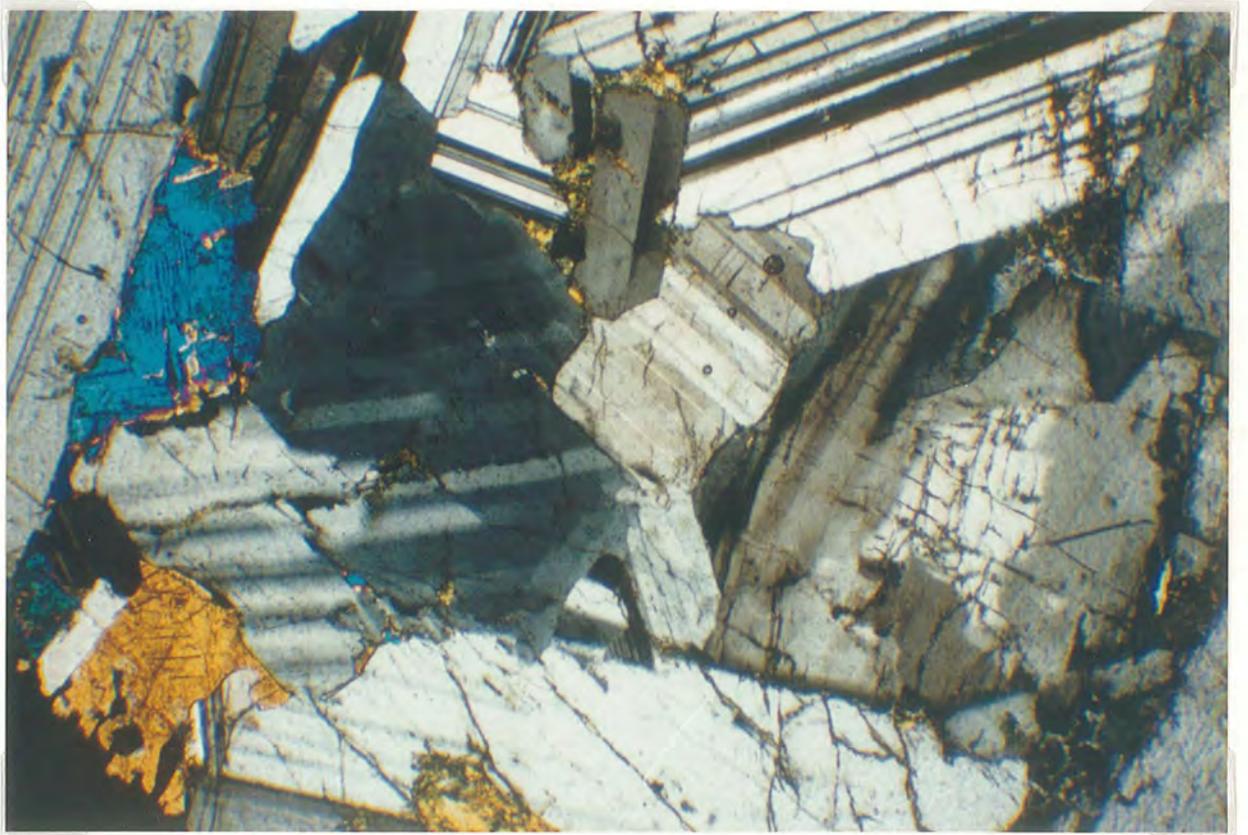
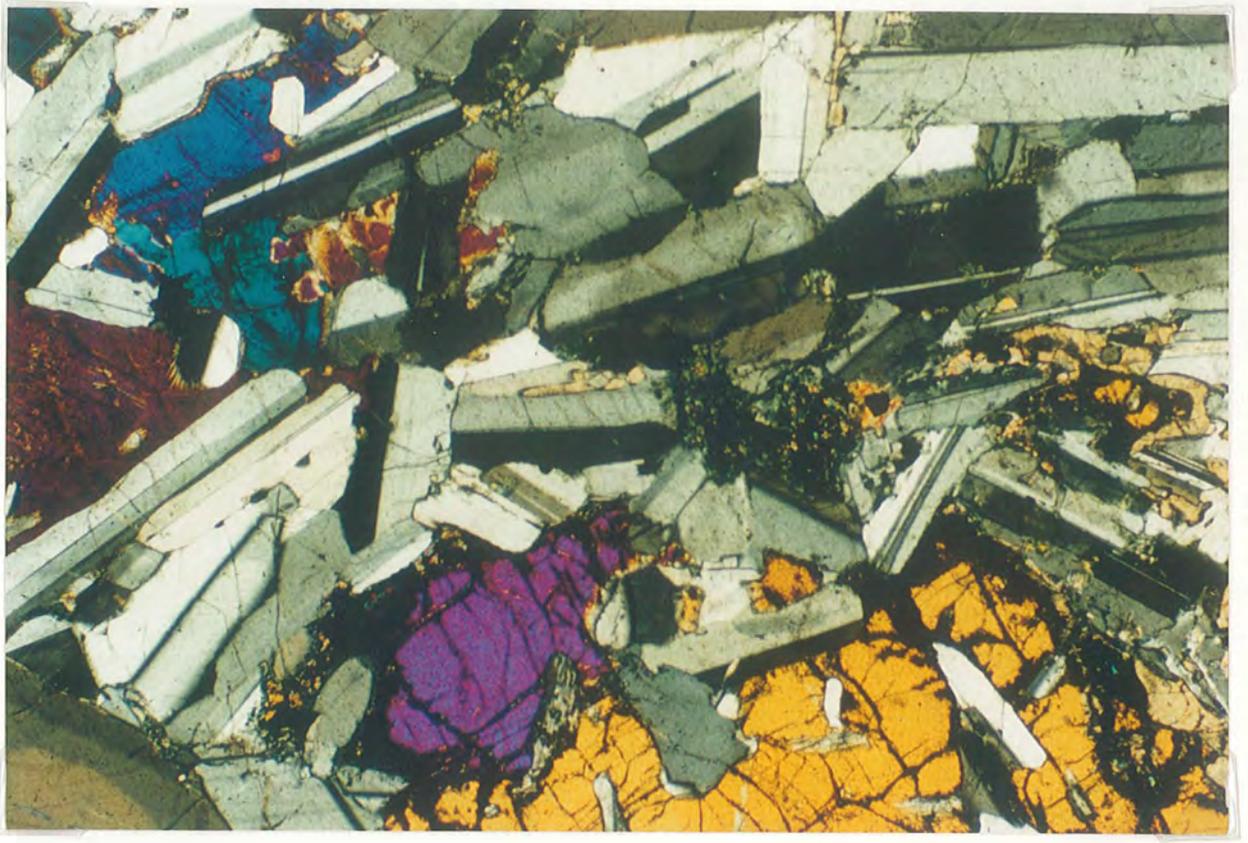
Other possible primary ( or at least, early - formed ) layers in the laminated sequence include granular anorthositic layers near its base, around 44007004, and some lamination - parallel augite - rich layers. The most pronounced modal variation in these rocks, which has in the past been considered to be part of the layered sequence ( Wells 1954 ), does however form part of the secondary pyroxenite layers ( section 3.4.3 ). In addition to these, the rocks contain small ( centimetre scale ) diffuse patches of coarser, isotropic olivine- and/or augite- rich rock which cross - cut the lamination and are particularly common close to the pyroxenites. These also seem to be of secondary or metasomatic origin.

Like the gabbronorites, the laminated gabbros in the sequence along the coast at Sanna contain xenoliths. The most common type is a fine - grained, finely banded and granular - textured anorthositic gabbro or troctolite, which also forms a lamination - parallel, intensely veined sheet up to 20cm thick in outcrops a few metres wide at the top of a crag at 44017001. Coarser isotropic anorthosite occurs as a series of lensoid xenoliths at 44057003. The lamination and layering in the adjacent rocks is symmetrically deflected around them. A secondary pyroxenite is also deflected beneath the largest xenolith, which may indicate that deformation and compaction around the xenolith continued until a relatively late stage, although there is no sign of particularly intense compression in the rocks above and below it.

Laminated fine - grained gabbros similar to the above rocks occur in a single outcrop, at 44106978, associated with the main occurrence in the Sanna area of a second class of pre - gabbronorite rocks: coarse anorthosite and anorthositic gabbro. This rock type forms a series of irregular, lobate and veined sheets in Grid Square 441697, ( see Fig. 3.25 ) with well exposed contacts against younger gabbronorites ( cf. Plate 3.81, above ), as well as a much larger but less easily defined ( because of weathering and staining of the outcrop ) anorthosite sheet around 440698, which may continue along

**Plate 3.83.** Plagioclase lamination in pre - gabbro laminated gabbro ( sample 318 ) deflected around olivine oikocryst at lower right - hand side of plate. Note that the oikocryst contains randomly orientated plagioclase laths, confirming that it pre - dates the development of the lamination. Crossed polarisers, width of field of view 3.5mm.

**Plate 3.84.** Pre - gabbro anorthosite, Sanna ( sample 287C4 ). Note well - developed internal straining evident in almost all the plagioclase crystals and the extensive corrosion of calcic phenocryst cores. Crossed polarisers width of field of view 3mm.

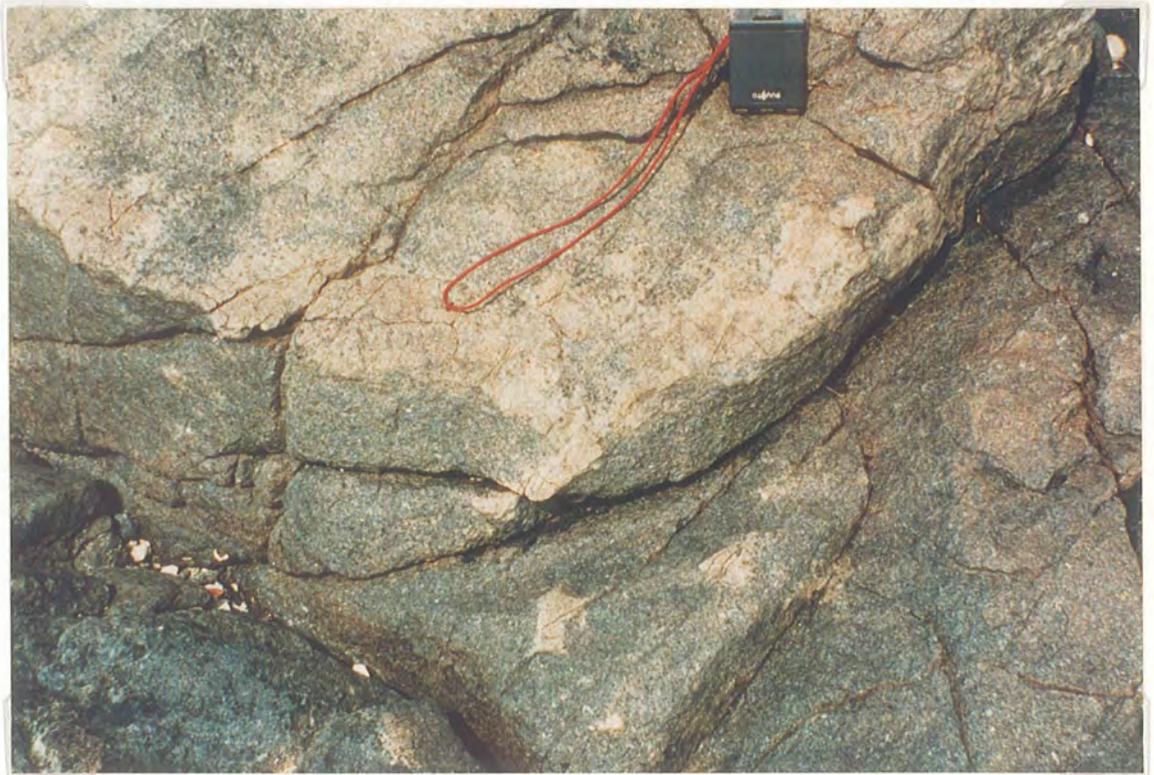


strike to a similar inclined sheet at the western end of Sanna Island. These rocks are characterised by a very high content of equant, subhedral to irregular plagioclase crystals 2 -20mm long, with intense internal straining, subgrain development, and corrosion of unzoned cores ( Plate 3.84 ). This deformation largely predates the augite oikocrysts which form much of the rest of the rock: these enclose strained feldspars and are themselves usually undeformed ( Plate 3.85 ). The plagioclase crystals in these oikocrysts are ragged and intensely corroded, implying partly replacive growth of these augites. Other undeformed interstitial poikilitic phases in these rocks are opaques and, in some, olivine rimmed by orthopyroxene. These rocks do not seem to be ultrafeldspathic equivalents of the gabbronorites: olivine and orthopyroxene are not as abundant in most, and bytownitic feldspar cores are rare. However, many of the larger feldspars are morphologically similar to the calcic porphyrocrysts in the gabbronorites. The sheets show signs of brecciation and disaggregation at their margins ( Plate 3.86 ), and evidence of partial melting and assimilation in the form of coarse anorthositic gabbro rims and backveins extending into the host gabbronorites, and are one potential source for the bytownite crystals in the gabbronorites. The numerous coarse anorthositic xenoliths which occur in some of the gabbronorite sheets may also be derived from pre - gabbronorite rocks within the Inner Series belonging to this group of rocks.

A final group of pre - gabbronorite rocks in the Sanna area outcrops in two localities, around 44506998 and on the northern side of a small peninsula just north of the contact with the Great Eucrite, around 446694. These are characterised by fine grained granoblastic and poikiloblastic textures, an absence of lamination, and exotic mineralogies. The northern locality consists of a series of shallowly south - dipping tabular and lensoid xenoliths and a larger irregular sheet, 2m thick and 15m long from east to west. They are enclosed in coarse porphyritic gabbronorite which sharply truncates banding within the granular rocks, except in a few places where a coarse isotropic pyroxene - rich rock forms a zone up to 20cm thick between the gabbronorites and granular anorthositic rocks within the xenoliths. The granular textured rocks are of two types: an anorthositic rock with wispy opaque-, augite- and, most unusually, apatite- rich bands and lenses as little as 1mm thick and a few mm long ( Plate 3.87 ); and a weakly banded augite - rich basic rock, which overlies the anorthosite. The augite - rich rock contains granular anorthosite bands and transgressive pods and veins of coarse - grained granular to poikilitic - textured augite - anorthite gabbro or eucrite. The augite is unusual in showing strong pleochroism in emerald greens and yellow, with a slight zonation to more intensely coloured rims, giving it a superficial resemblance to aegirine ( see, however, analyses of these pyroxenes in Appendix

**Plate 3.85.** Undeformed, post - deformational augite oikocryst enclosing intensely strained and corroded plagioclase crystals. Pre - gabbroanorthosite, Sanna ( sample 287B1 ). Crossed polarisers, width of field of view 3mm.

**Plate 3.86.** Disrupted remnants of anorthosite sheet in gabbroanorthosite, Sanna. Field location 287A, grid reference 44186978 approx.

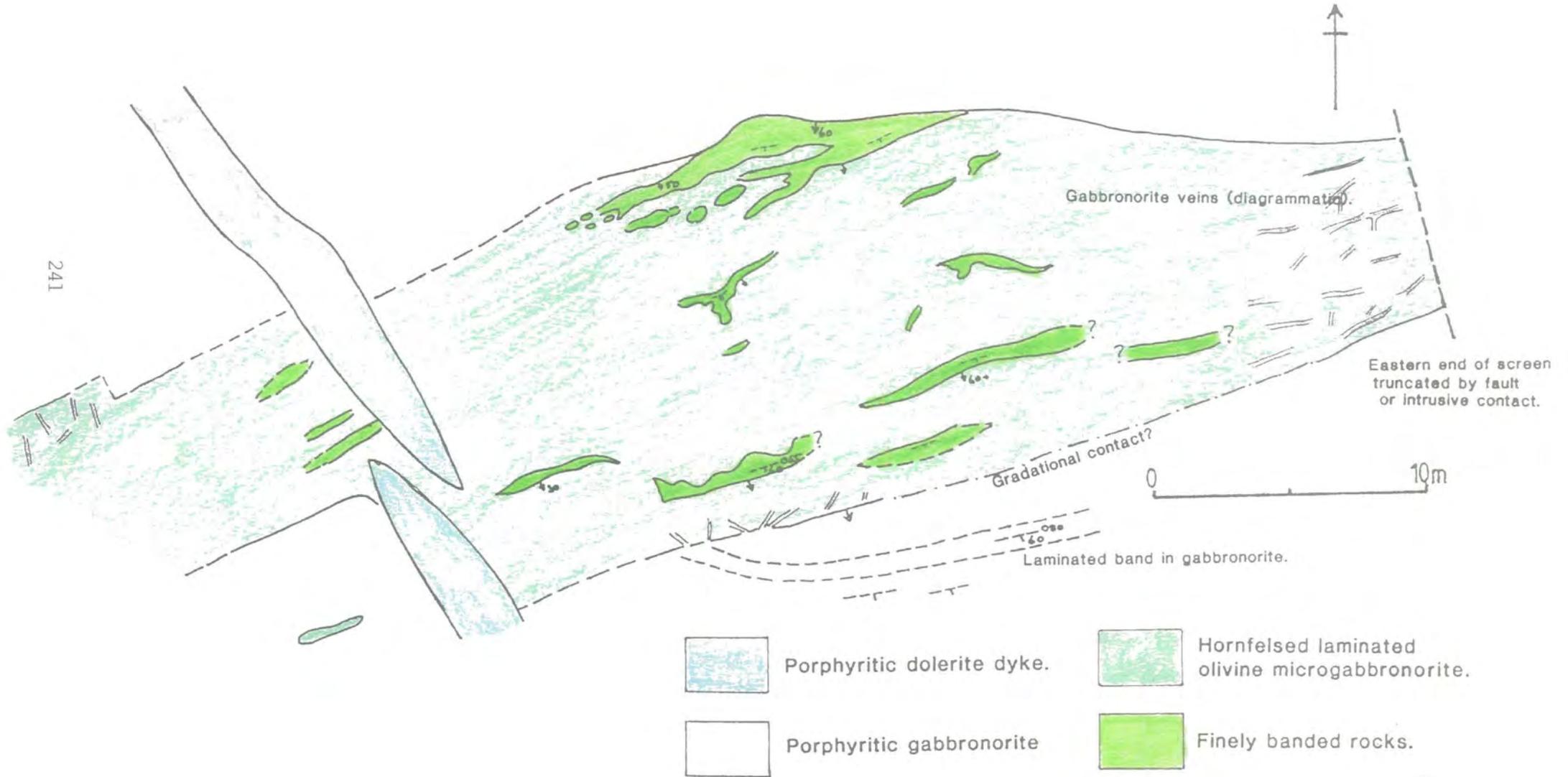


3, sample 312B3 ). In addition the feldspars in these rocks are extremely calcic ( calcic bytownite and anorthite; see Chapter 6, section 6.3 ).

The southern locality is even more complex. It consists of a series of very large, steeply SSE - dipping blocks, offset from one another and veined by the host gabbroites. The largest of these blocks, some 40m long and 15m wide at maximum, was mapped in detail ( Fig. 3.26 ). The other blocks, further to the west, are not as well exposed but conform to the same general pattern. Each is mainly composed of homogenous fine - grained laminated granular troctolitic microgabbroites which contain discontinuous irregular strips of extremely finely layered feldspar - rich rock. The latter contains laterally continuous bands, 0.1 to 5mm thick, of augite and plagioclase with some opaque grains. These bands are commonly intensely and complexly folded, mainly around NNW - SSE trending subvertical sinuous fold axes which plunge very steeply to the south ( Plate 3.88 ). The style of folding, with folded fold axes, sharp variations in the geometry of the fold limbs along the length of the fold axes and, in places the formation of rootless and intrafolial folds, bears a strong resemblance to soft sediment deformation. In thin section, the rocks have a granular texture and show wide variation in grain size between adjacent layers, independent of composition ( Plate 3.89 ), and rare porphyrocrysts are present. The grain size variation suggests that the rock has undergone recrystallisation and textural re - equilibration after a period of intense, strongly heterogeneous deformation and grain size reduction to different grain sizes in different areas of the rock. Rare post - deformational augite poikiloblasts occur in some bands within the rock, as do sub - poikiloblastic to interstitial sphene and epidote crystals: the latter may, however date from later, lower - grade metamorphism since they only occur in these outcrops, which are less than 100m from the contact with the later Great Euclite intrusion. The banded rocks are cross - cut by coarse, heterogeneous augite - anorthite gabbro or euclite veins which are themselves sometimes folded ( Plate 3.90 ). These veins do not cut the laminated microgabbroic rocks which form the immediate host to the finely - banded rocks. Both the banded rocks and the veins contain very calcic plagioclase and emerald green to yellowish pleochroic augite similar to those in the banded augite - rich granulites around 44006998. Pyrite is present as prominent large grains and flakes but may be secondary in origin.

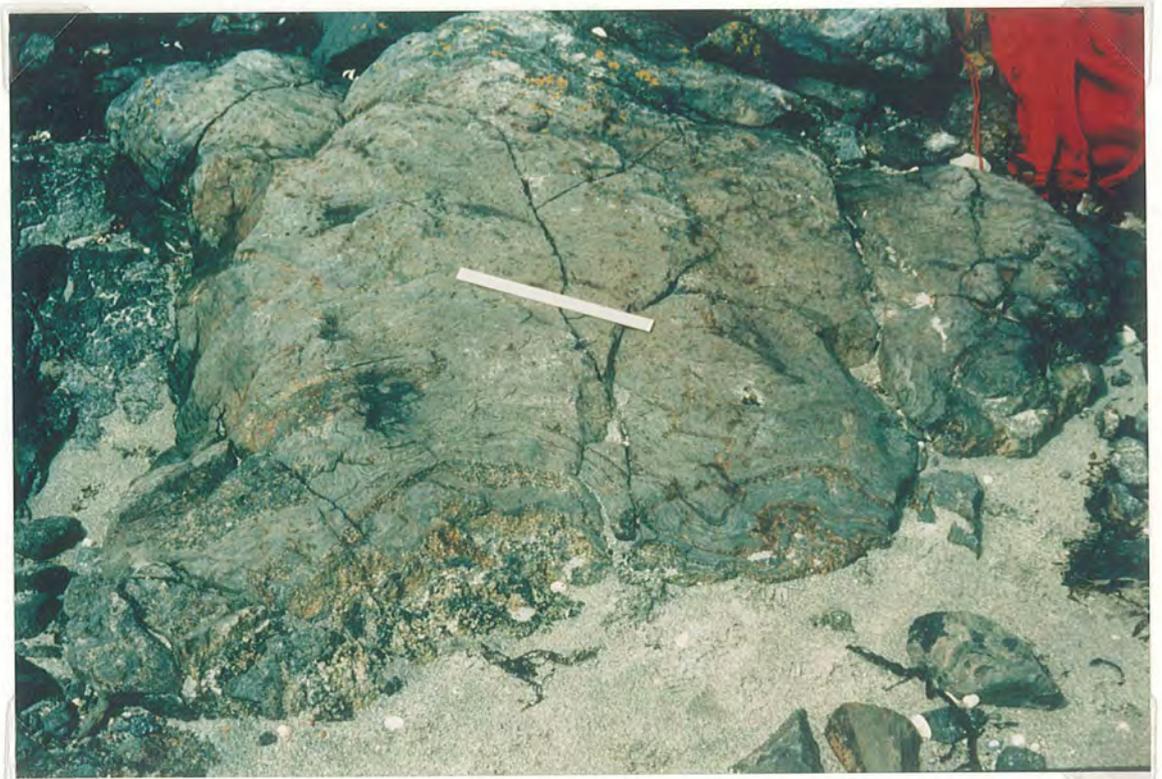
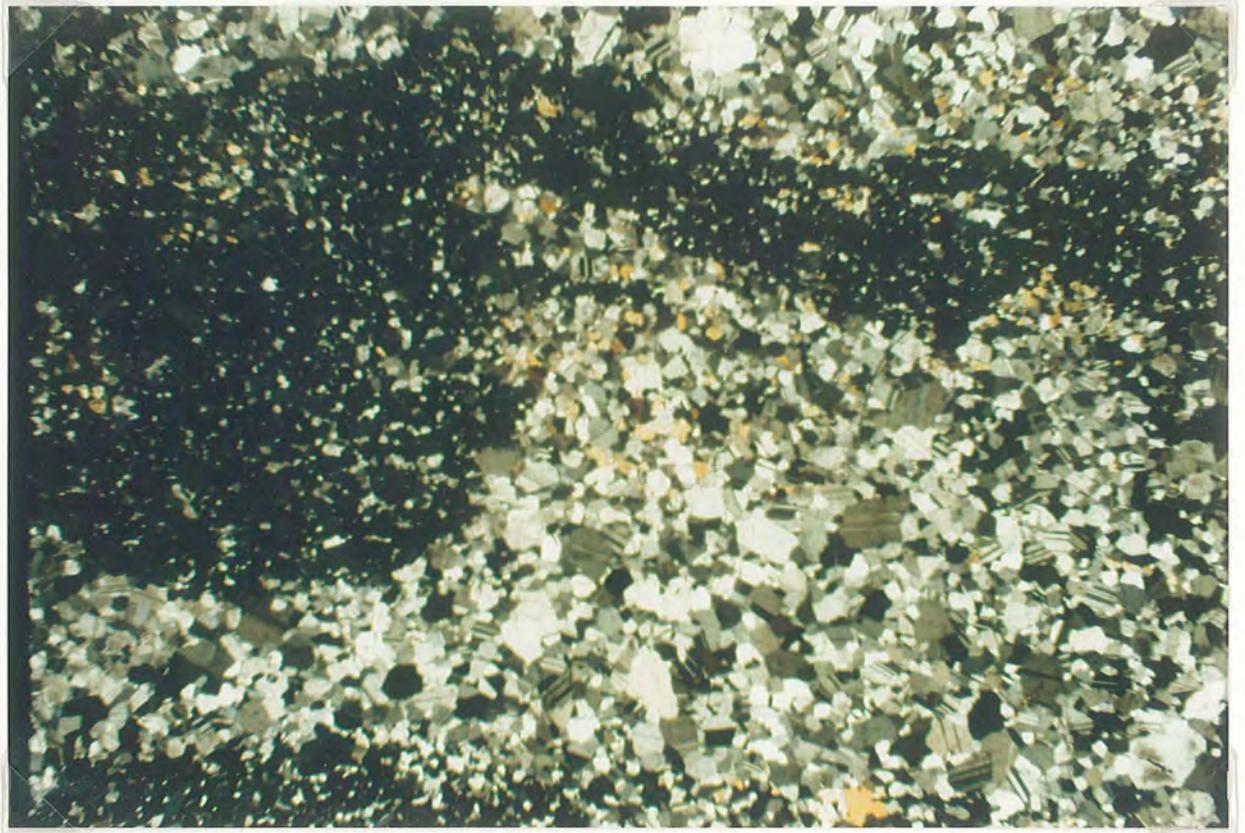
The host rock to the finely banded rocks is completely unrelated, being a troctolitic, orthopyroxene - bearing microgabbro with a few thin olivine - rich bands ( see Plate 3.113, section 3.4.6 ). A weak

FIG. 3.26. MAP OF XENOLITHIC SCREEN IN GABBRONORITES, SANNA BAY.



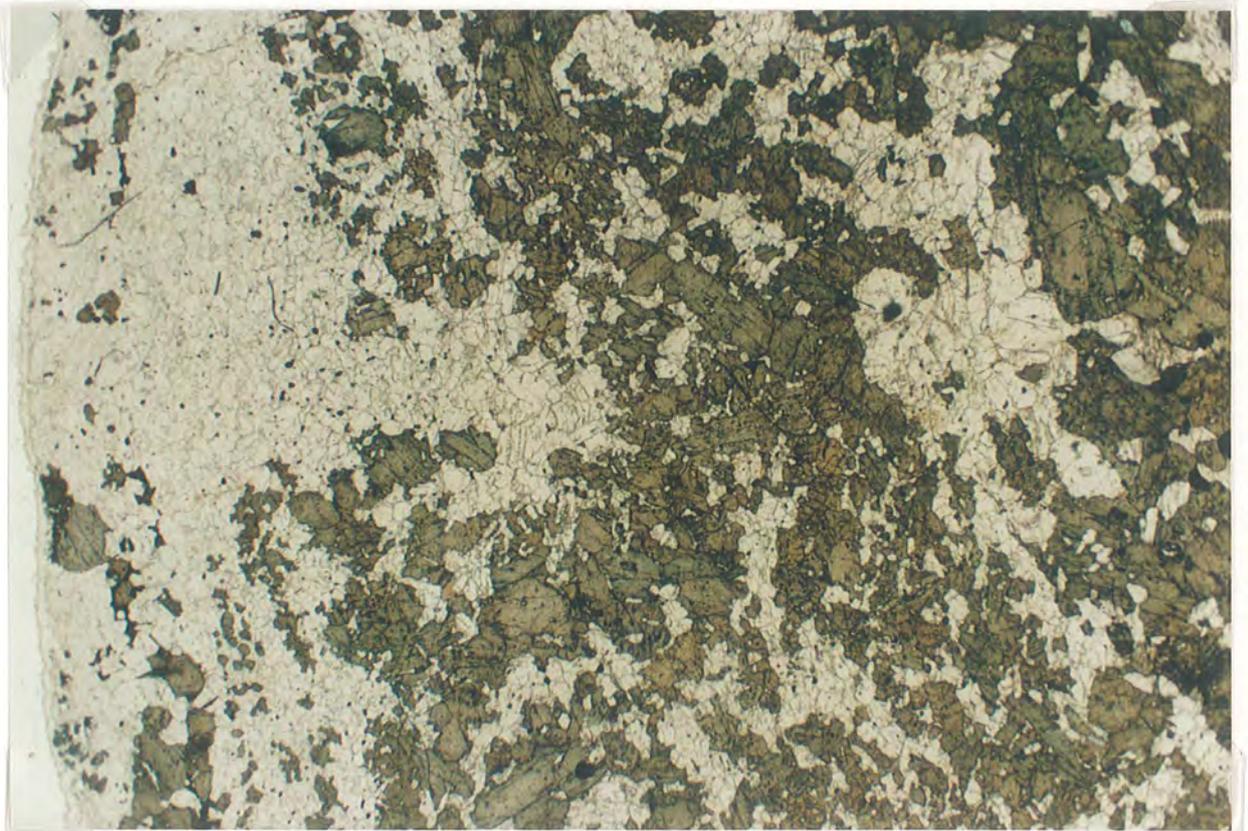
**Plate 3.87.** Near - isotropic apatite - rich lenses and pods in anorthositic rock ( pre - gabbronorite granular anorthite - rich suite ). Sample 36/2. Crossed polarisers, width of field of view 14mm.

**Plate 3.88.** Banded granular augite - anorthite gabbro or eucrite ( feldspar - rich ). Xenolith within xenolithic screen, field location 97/1 ( see Fig. 3.26 ). The folding of the banded rocks pre - dates the pegmatoid veins in the foreground, which themselves predate the host laminated microgabbronorite. Grid reference 44176944.

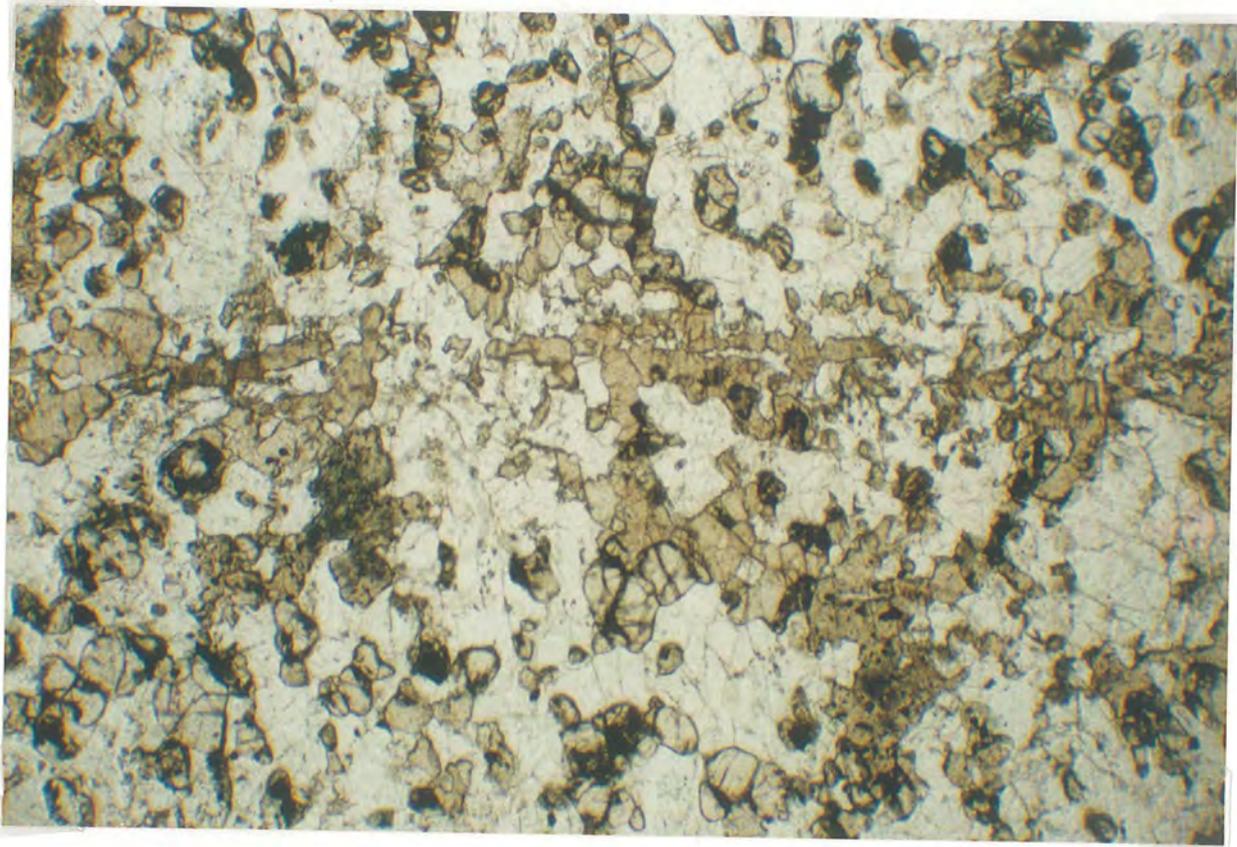


**Plate 3.89.** Banded granular augite - anorthite gabbro. Note large ~~sphene~~ poikiloblasts. Sample X1. Plane - polarised light, width of field of view 14mm.

**Plate 3.90.** Augite - anorthite vein cutting banded granular augite - anorthite gabbro. Sample 97/5. Plane - polarised light, width of field of view 14mm.



**Plate 3.91.** Laminated granular olivine microgabbro ( sample 97/1 ). Cut by annealed orthopyroxene - rich hydrothermal vein. Plane - polarised light, width of field of view 3.5mm.



lamination defined by preferred lattice orientation of feldspars is present but the morphological lamination is largely obscured by subsequent development of a granular to poikiloblastic texture ( Plate 3.91 ).

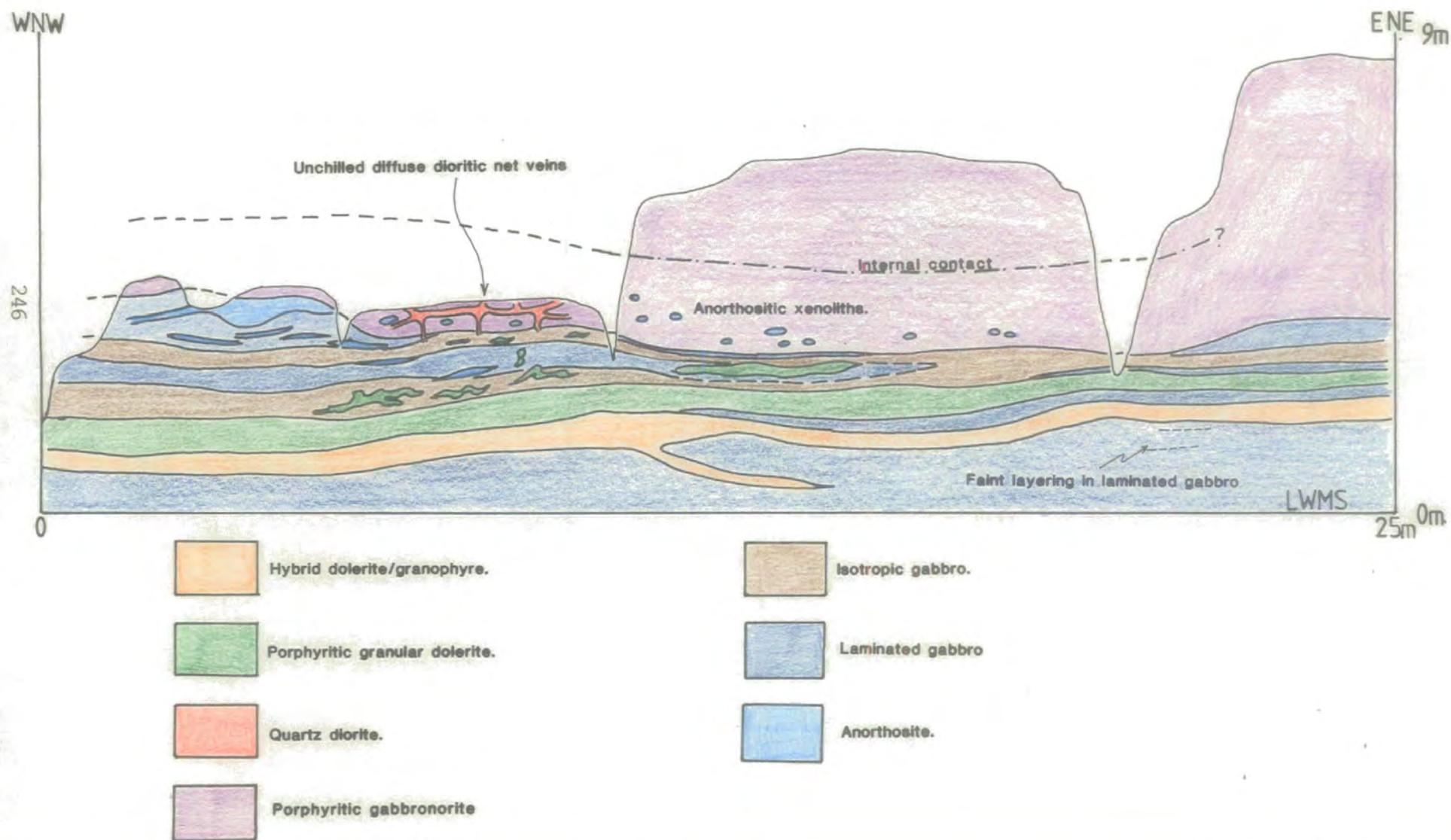
The intense deformation in these rocks does not affect the adjacent gabbro norites, nor the angular gabbro net veins which cut parts of the outcrop ( Fig. 3.26, above ) and must, along with the augite - anorthite gabbro veins, pre - date the intrusion of the rocks by the gabbro norites. The steeply - south dipping contacts are sub - parallel to the banding and lamination in the block ( which may be coincidental ) and also to weak lamination in the gabbro norites to the south, although sharply discordant NNW - SSE trending contacts are also present ( Fig. 3.26, especially at the western end of this map ). The gabbro norites to north and south are subtly different, the latter being coarser grained and more feldspar - rich. These features indicate that the intervening older rocks form a screen between separate gabbro norite intrusions, like those formed by the laminated gabbros to the north of Sanna Bay. The screen seems to represent a remnant of a laminated intrusion which itself contained the disrupted remnants of a series of finely banded, deformed and subsequently veined cumulate rocks of anomalous composition.

#### **3.4.2.2. Pre - gabbro norite rocks at An Acairseid.**

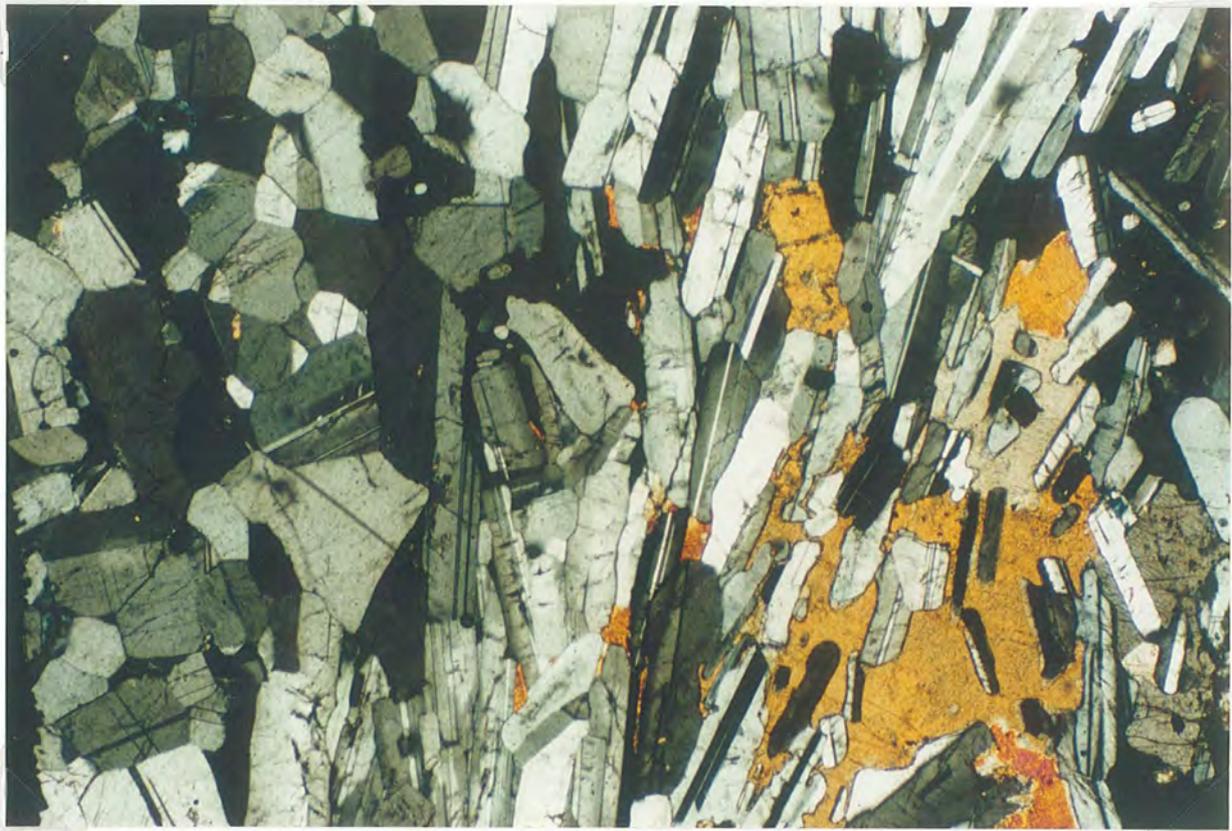
Laminated fine - grained and coarser isotropic gabbros occur beneath a relatively fine - grained gabbro norite sheet with prominent plagioclase porphyrocrysts in coastal outcrops around 43536333 ( see Map 4 ). The visible sequence above the low tide mark is about 6m thick in total, is intruded by a number of beerbachitic dolerite sheets and veins ( see section 3.4.4 ) and a later dolerite/granophyre mixed - magma intrusion. The lamination and layering within the sequence dip  $8 - 12^\circ$  ( strike  $110^\circ$  ), a slightly shallower angle than the dips of the contact with the gabbro norite and of the main dolerite sheet within the laminated sequence. The contact with the gabbro norite also cuts up- and down- section through the sequence in a strike - parallel direction ( Fig. 3.27 ).

The pre - gabbro norite rocks in this sequence are characterised by more pronounced modal and textural variations than occur in the laminated gabbros at Sanna which are their nearest equivalents. Three main rock types are present in laterally discontinuous layers: anorthosite, ophitic to granular laminated gabbros and isotropic oxide - rich quartz gabbros. The entire sequence is intensely altered

FIG. 3.27 CROSS-SECTION SHOWING PRE-GABBRONORITE ROCKS AT FIELD LOCATION 275, AN ACAIRSEID (43536333 approximately: see Map 4).



**Plate 3.92.** Very strongly laminated augite - anorthosite rim around adcumulate - textured anorthosite xenolith in gabbro. Sample 275/1A. Crossed polarisers, width of field of view 3.5mm.



( section 3.4.6 ), which complicates the task of interpreting these rocks.

The anorthosites, which occur near the top of the sequence in discontinuous layers and pods, are fine - grained rocks with a polygonal adcumulus texture, unlike the polymineralic rocks which retain primary igneous textures and are incompletely texturally equilibrated. This is consistent with the hypothesis that subsolidus textural equilibration is fastest in monomineralic rocks ( Hunter 1987 ), and does not necessarily imply that the anorthosites are a separately - formed group of rocks. Xenoliths of adcumulate anorthosite, with peculiar coatings of exceptionally well - laminated feldspar - rich rock with very elongate feldspar laths orientated tangentially to the xenolith core ( Plate 3.92 ) and poikilitic - interstitial augite and opaques, occur in the overlying gabbro and are probably derived from the adjacent anorthosites.

The laminated gabbros are very similar to those at Sanna, in both texture and mineral assemblage. The lamination again appears to be a deformational structure, on the same evidence as was described above for the rocks at Sanna. The isotropic gabbros, however, have no equivalents at Sanna. They lack olivine altogether and contain interstitial quartz and quartz - alkali feldspar intergrowths in addition to small amounts of orthopyroxene. They have a high modal augite content ( about 50% ) but plagioclase occurs as euhedral laths rather than as corroded and interstitial grains as is the case in the secondary pyroxenites ( section 3.4.3 ). The feldspar is a sodic labradorite with a restricted compositional range and calcic cores are absent. Opaques are abundant and coarse - grained but irregular to interstitial, and may be partly secondary in origin. These isotropic gabbros have very well - preserved primary igneous textures and sharp contacts, down to the scale of individual grains, with the laminated rocks. At the top of the layered sequence in Fig. 3.27 they enclose elongate lenses of laminated gabbro. These features suggest that the isotropic gabbros may have been intruded into the laminated gabbros, which would also explain the differences in mineralogy and mineral compositions. The absence of similar isotropic gabbros intruded into the overlying gabbros suggests, however, that both the laminated and the isotropic gabbros predate the gabbros.

#### **3.4.2.3. Opaque - Rich and Laminated rocks on and to the south of Beinn na Seilg.**

Pre - gabbro rocks occupy a large part of the southern and western sides of the summit ridge of Beinn na Seilg above the 300m contour. They also occur in a number of isolated outcrops on

the south side of the hill and in the valley below, over a vertical distance of almost 200m ( see Map 1 ). These rocks typically occur in irregular sheet - like and tabular bodies with northward dips, towards the centre of the pluton, on both upper and lower surfaces. They are cut by both the gabbronorites and, in many places, by still later beerbachitic dolerite sheets, which between them have produced much disruption of these rocks. As observed by Skelhorn & Elwell ( 1971 ), the pre - gabbronorite sheets and the layering within them show an overall northward increase in dip from 10 - 15° northwards in a sheet extending from 45406357 to 45406353, to as much as 60°+ in outcrops on the summit ridge of Beinn na Seilg. However, the dips in this latter area are very varied and mostly range from 20° to 40°. The actual significance of many of these dip values are discussed in section 3.4.7, below. Although the upper and lower surfaces of these tabular bodies are broadly subconformable to lamination and layering within them, they are cross - cut by numerous irregular gabbroic veins. Furthermore, where the ends of the bodies are actually exposed, these internal structures are mostly truncated at high angles by the enclosing gabbronorites, confirming the xenolithic or screen - like character of these rocks.

The pre - gabbronorite rocks in the area fall into two groups. Most of the outcrops around the summit of Beinn na Seilg are of the first type. These are rather granular microgabbros with a well - defined feldspar lamination but containing mostly granular, rather than poikilitic, augite, orthopyroxene and opaques. They lack olivine and mostly have an unusually high content ( up to 15% ) of magnetite in textural equilibrium with the rest of the rock and therefore of apparently igneous origin, in contrast to the secondary, fine - grained opaques which are also present ( see section 3.4.6 ). Weak layering is present in some of these rocks, as it is in a few outcrops of the weakly laminated variants of the host gabbronorites which occur in this area. A further complication is that some of the laminated gabbros contain rare angular xenoliths of even finer - grained, more strongly laminated and more granulitic - textured oxide - rich microgabbros. However, apart from their more fractionated and Fe<sup>3+</sup> - rich mineralogies and the opaque - rich xenoliths, these rocks do bear a general resemblance to the laminated gabbros at Sanna and An Acairseid.

The second group occurs in a large sheet on the eastern side of the gully running north from the head of the Dubh Chreag gorge ( outcrop from 45406357 to 45406353 ), in two small outcrops around 45716415 on Beinn na Seilg and as a xenolith in an adjacent beerbachitic dolerite sheet ( 45626415 ). These are very distinctive, mainly plagioclase - magnetite rocks with well - defined but frequently

irregular and discontinuous bands of magnetite, anorthosite, troctolite and magnetite - troctolite in addition to the volumetrically dominant magnetite anorthosite. Augite is only rarely present in these rocks and non - metamorphic orthopyroxene is entirely absent. The banding is on a millimetre to centimetre scale and typically undulating, folded or irregular ( Plate 3.93 A and B ). Very fine - grained adcumulate and heteradcumulate textures, with very large poikilitic unstrained olivines ( Plate 3.94 ), are typical, although these are often masked by metamorphic oxidation of the rocks ( see section 3.4.6 ). A well - defined plagioclase lamination is present, particularly in oxide - poor layers.

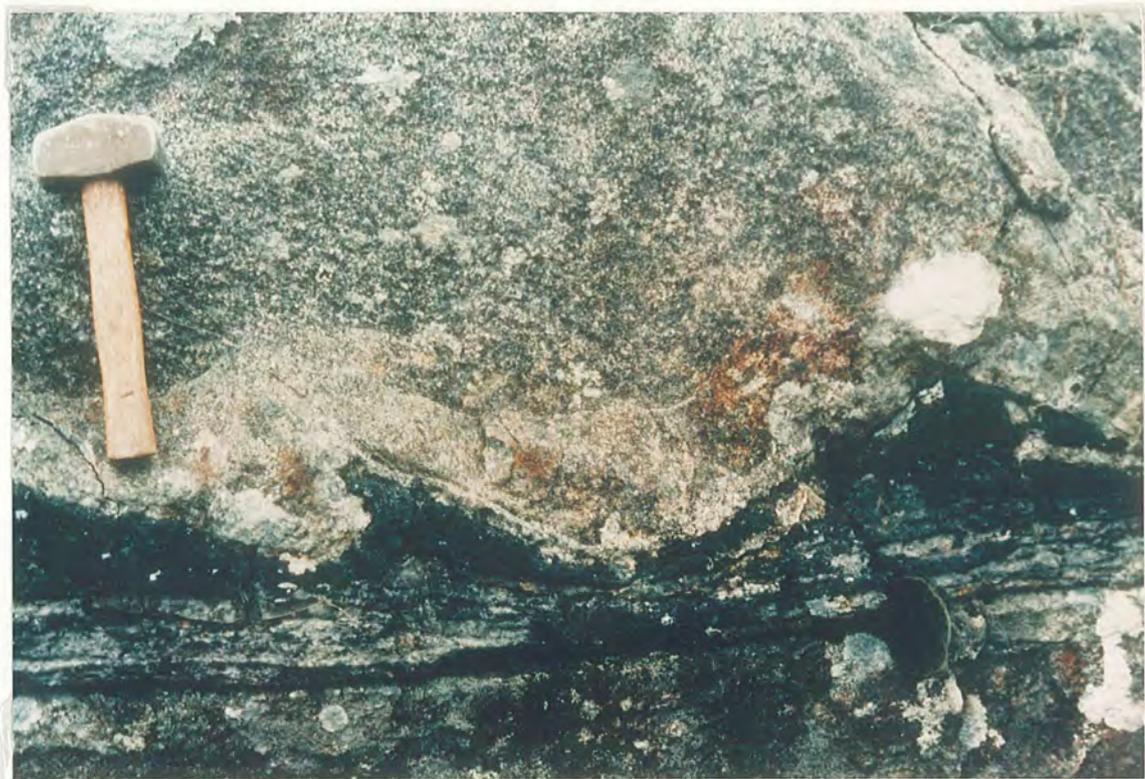
#### **3.4.2.4. Other occurrences of pre - gabbonorite rocks in the Inner Series; and some comments on their overall distribution.**

A few isolated, and mainly small, outcrops of pre - gabbonorite rocks occur in other parts of the Inner Series. Layered, fine grained polygonal anorthosite / poikiloblastic magnetite troctolite xenoliths, which are petrographically similar to those in the Beinn na Seilg area, occur south - east of Ardnamurchan point at 41836704 and on the eastern side of Stacan Dubha, to the east of Beinn na Seilg. An isolated outcrop of layered rocks with their upper surface cross - cut by gabbonorite occurs in a cliff north of An Acairseid, at 437639. Finally, coarse oikocrystic laminated olivine leucogabbro outcrops at 42606680, east of Ardnamurchan Point, in a series of thin ( 10 - 50cm thick ) layers alternating with gabbonorites. The leucogabbro layers are well - defined and laterally continuous but have lobate margins and interfinger with the gabbonorites in places ( Plate 3.95 ). This type of lobate margin is also developed at the contacts between gabbonorite and anorthositic gabbros in the outcrops at Sanna, and these rocks may be equivalent to the latter, although they are more olivine - rich. Geochemical evidence relevant to the relationships of these various anorthositic rocks is considered in section 6.3.

The pre - gabbonorite rocks in the Inner Series are of considerable potential interest because age relationships of these rocks to the Marginal Border Group are such that it is possible that some of them may be cumulates corresponding to the liquid compositions represented by the more - or - less quenched basic rocks in the Marginal Border Group. Of the rocks described in this subsection the laminated, weakly banded orthopyroxene - bearing gabbros are the most plausible candidates. This is because they have the requisite tholeiitic mineralogy and are widely distributed through the Inner

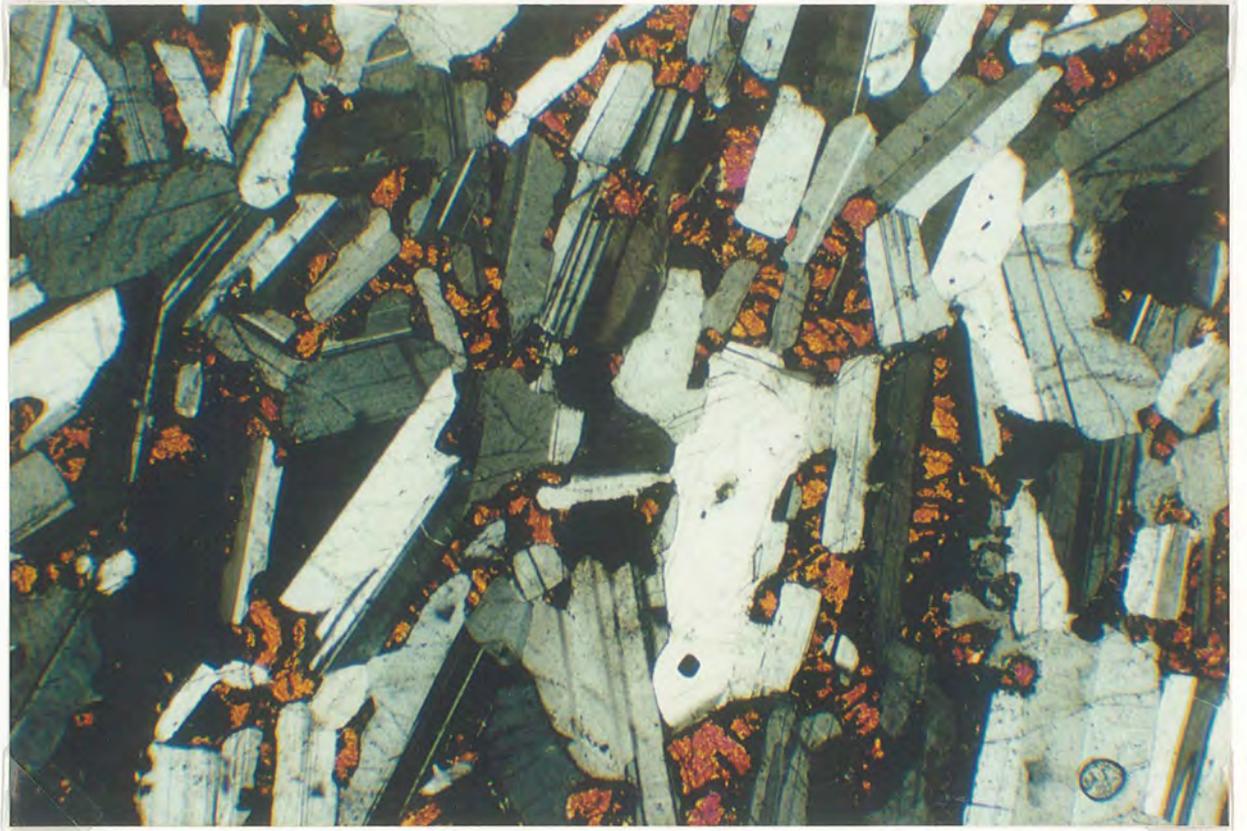
**Plate 3.93A.** Banded granular magnetite - anorthosite rock, forming xenolithic sheet in gabbro norite south west of Beinn na Seilg. Field location 166B, grid reference 45406355.

**Plate 3.93B.** Deformed banded magnetite - anorthosite rock, xenolith on Beinn na Seilg. Field location 284, grid reference 45716415. It should be noted that the sample from this locality ( Plates 3.94 and 3.111 ) were collected from an adjacent xenolith with the same mineralogy and banded structure, but less deformed.



**Plate 3.94.** Laminated granular anorthosite ( sample 284 ) containing large interstitial - poikilitic undeformed ( post - lamination ? ) olivine grains. These show oxidation to magnetite alone ( Plate 3.111 ).

**Plate 3.95.** Lobate sheet of olivine anorthosite with prominent large olivine oikocrysts, enclosed in laminated gabbronorite. Field location 137 ( grid reference 42606680 ).



Series, as would be expected for cumulates formed on the floor of the MBG magma chamber, which must have overlapped at least most of the region now formed, at the surface, by the Inner Series.

The presence of other, highly distinctive cumulate ( or at least, cumulate - like ) lithologies as screens and xenoliths within the Inner Series carries the implication that a number of pre - Inner Series intrusions besides the Marginal Border group magma chamber must have been present in the region broken up by the intrusion of the gabbroites of the Inner Series. The troctolitic and plagioclase - magnetite rocks on the one hand, and the augite - anorthite gabbros on the other, cannot have crystallised from the same, or even compositionally similar, batches of magma ( see Chapter 6 ). Given the evidence for central uplift at the time of emplacement of the Inner Series, it seems likely that these would have been emplaced originally well below the present - day level of exposure.

### **3.4.3 Secondary or Replacive Olivine Pyroxenites.**

Apart from the various pre - gabbroite rocks, the most pronounced modal variation within the Inner Series is seen in a number of discrete, subconformable feldspathic olivine pyroxenite bodies, by far the best examples of which are found in and around the outcrops of laminated gabbros to the north of Sanna Bay ( Fig. 3.25; also Figs. 3.23 and 3.24 ). The pyroxenites in this area form two thick ( up to .7m ) and numerous, mainly much thinner layers or layering - parallel veins. The thicker layers in particular are laterally extensive over distances of tens of metres, although they do die out gradually both along strike and down - dip. Pyroxenites of this type occur in both the laminated gabbros and in the adjacent gabbroites ( see Fig. 3.23 ).

The tops of two thickest pyroxenites are respectively 1m and about 3m below the base of the lower gabbroite sheet in Fig. 3.23 where it outcrops ( Grid Ref. 44026997 ) in the line of section on which this log was based. They are, however, markedly transgressive with respect to the base of this sheet. The Upper Main Pyroxenite ( UMP ) cuts up - section to the north and cuts the margin of the gabbroite, outcropping within it north of about 44077002, whilst the Lower Main Pyroxenite ( LMP ) also cuts up towards the gabbroite but within the exposed area remains within the laminated gabbros. Furthermore, although the LMP continues out to sea to the west the UMP dies out as a series of sub - parallel veins around 43997000, clearly visible in a NNW -

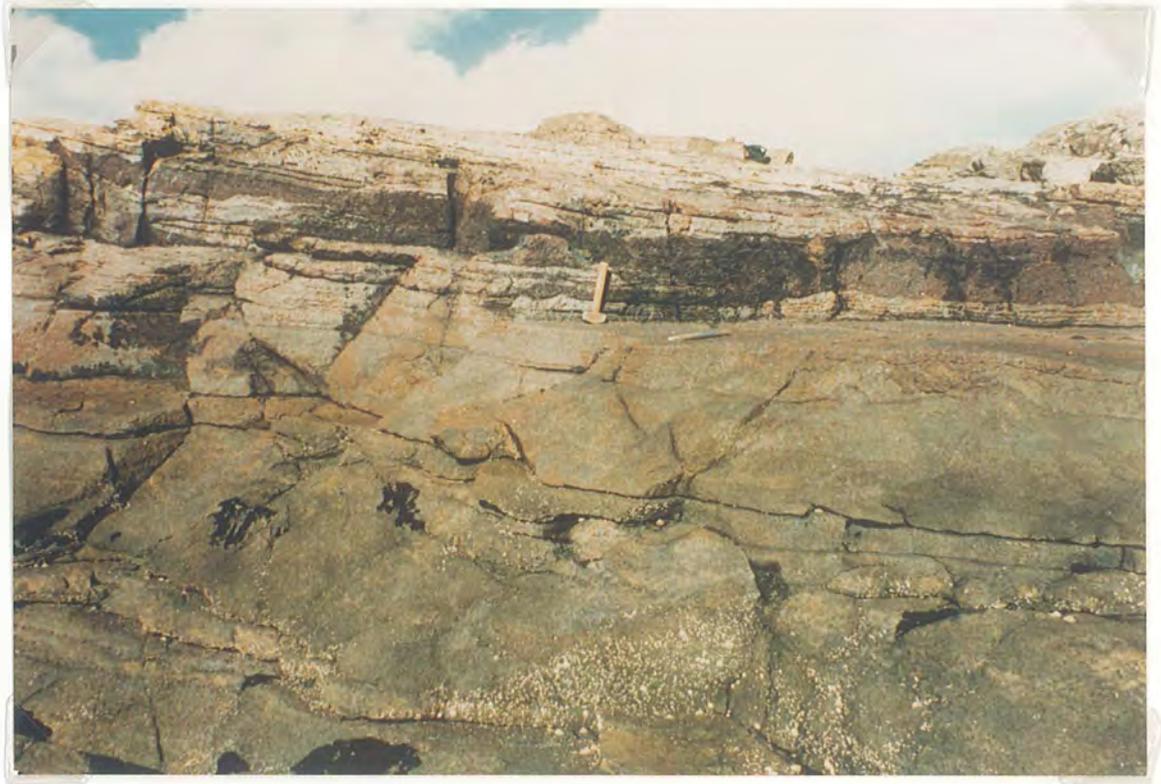
SSE trending gully wall ( Plate 3.96 ). These veins cross - cut but do not deform layering in the host laminated gabbros, implying that the pyroxenites formed by a metasomatic or ( at least approximately ) volume - for - volume replacement process. At the top of the sheet in this outcrop, and at the tips of the veins, the host gabbro contains coarse isotropic pyroxene - rich horizons ( Plate 3.97 ) which die out with distance from the pyroxenite. This suggests that pyroxenite formation was a process which exploited particular horizons in the laminated gabbro. To the east, the bases of both the UMP and the LMP are sharp and almost planar, but the tops are highly uneven, with lobes and replacive veins cutting the host gabbro ( Plate 3.98 ). Above both main pyroxenites there is a zone, of varying thickness, of coarse isotropic augite - rich gabbro, which itself has a distinct irregular contact against the overlying laminated gabbros. Pyroxene - rich gabbro 'fingers' ( Robins 1982 ) penetrating the laminated gabbros are, like the main pyroxenites, associated with preferential replacement of particular horizons in the laminated gabbros ( Fig. 3.28 and Plate 3.99 ). In contrast, where the UMP cuts isotropic gabbronorite the contact between the two is sharp and cusped, without preferential replacement of particular parts of the gabbronorite in this case. This presumably reflects the greater homogeneity of the gabbronorite in some property, such as composition or grain distribution, which affects the process of pyroxenite metasomatism in these rocks ( see Chapter 6 ).

Strong petrographic evidence for the metasomatic origin of the pyroxenites occurs in some of the relatively feldspar - rich pyroxenite or melagabbro layers which occur in the gabbronorites ( see Fig. 3.23 ) and also in those parts of the UMP which cut the gabbronorites. Although most of the feldspars in these are unzoned ragged interstitial laths and sub - poikilitic grains, a small proportion of the larger grains contain rounded calcic bytownite cores ( Plates 3.100A and B ). These are not present in pyroxenites and pyroxene - rich melagabbros in the laminated rocks, including that part of the UMP which occurs in the latter. This pattern of distribution of the porphyrocrysts suggests that they represent relict refractory porphyrocrysts from the gabbronorite replaced by the pyroxenitic rocks.

The interstitial character of the feldspar is much more pronounced in the LMP and UMP than in the smaller veins. Sections from the cores of both main pyroxenites contain very large, irregular olivine and orthopyroxene grains up to 2cm long, poikilitic to equant euhedral augite and orthopyroxene grains, and rare interstitial to poikilitic plagioclase and opaques ( Plate 3.101 ). At the upper

**Plate 3.96.** Western extremity of the Upper Main Pyroxenite, where it dies out into a series of sub - parallel thin sheets which replace particular horizons in the host laminated gabbro. Field location 316, grid reference 44017000 approx.

**Plate 3.97.** Detail from Plate 3.96 showing coarse isotropic horizons in the host gabbros thickening and coarsening towards the pyroxenite. This gradation suggests that the metasomatic process leading to pyroxenite formation was a gradual one which exploited particular horizons in the host rocks.



**Plate 3.98.** Lobate irregular upper margin of the Upper Main Pyroxenite where it is replacing laminated gabbros, field location 316. The rock immediately above the pyroxenite in this view ( photographed at a point about 5m east of that in Plate 3.96 ) is an isotropic coarse - grained gabbro formed by recrystallisation of the host laminated gabbro ( see Plate 3.99 ).

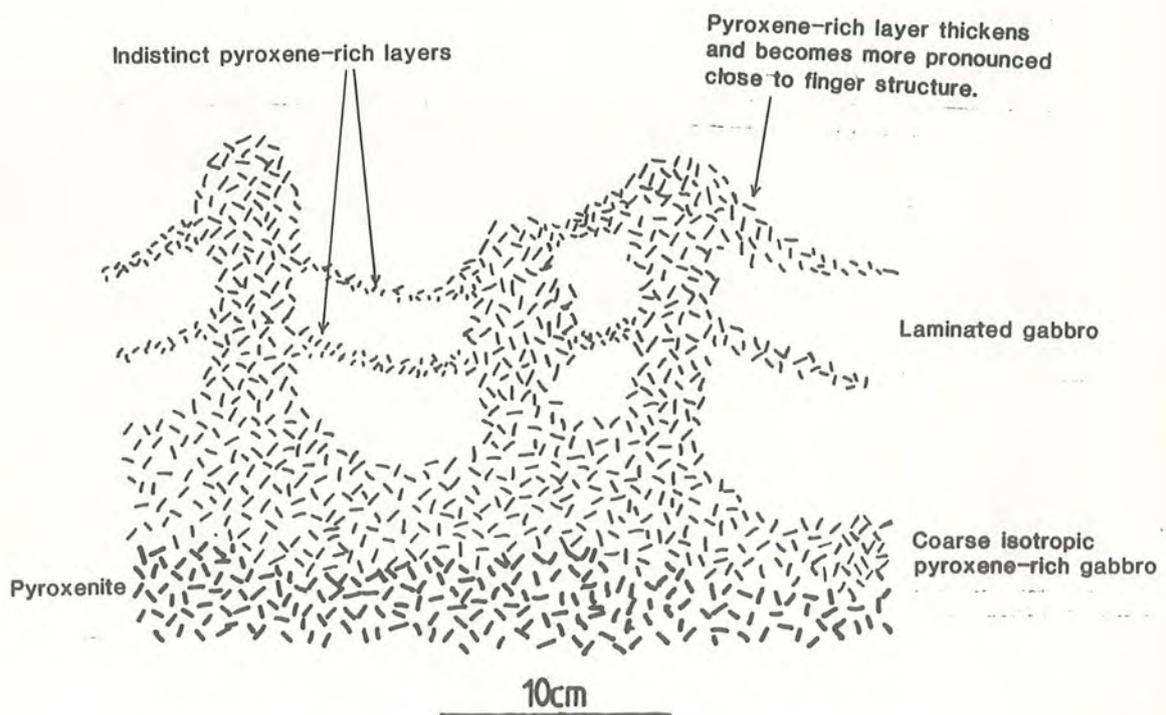


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Plate 3.99. Vertical rock surface showing "fingers" of pyroxene-rich gabbro cutting laminated gabbro, above Lower Main Pyroxenite, Sanna.

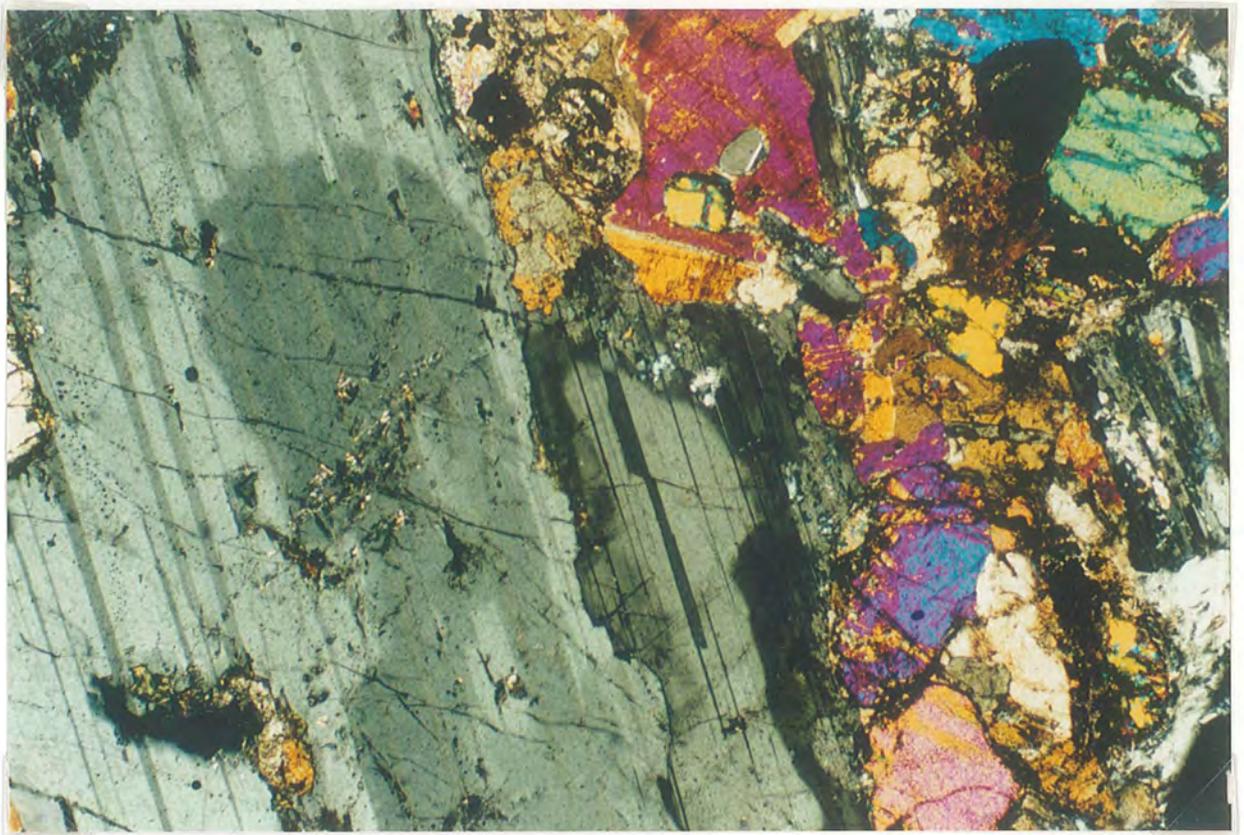


Fig. 3.28. Sketch of "fingers" cutting laminated gabbro: same outcrop as that in Plate 3.99.

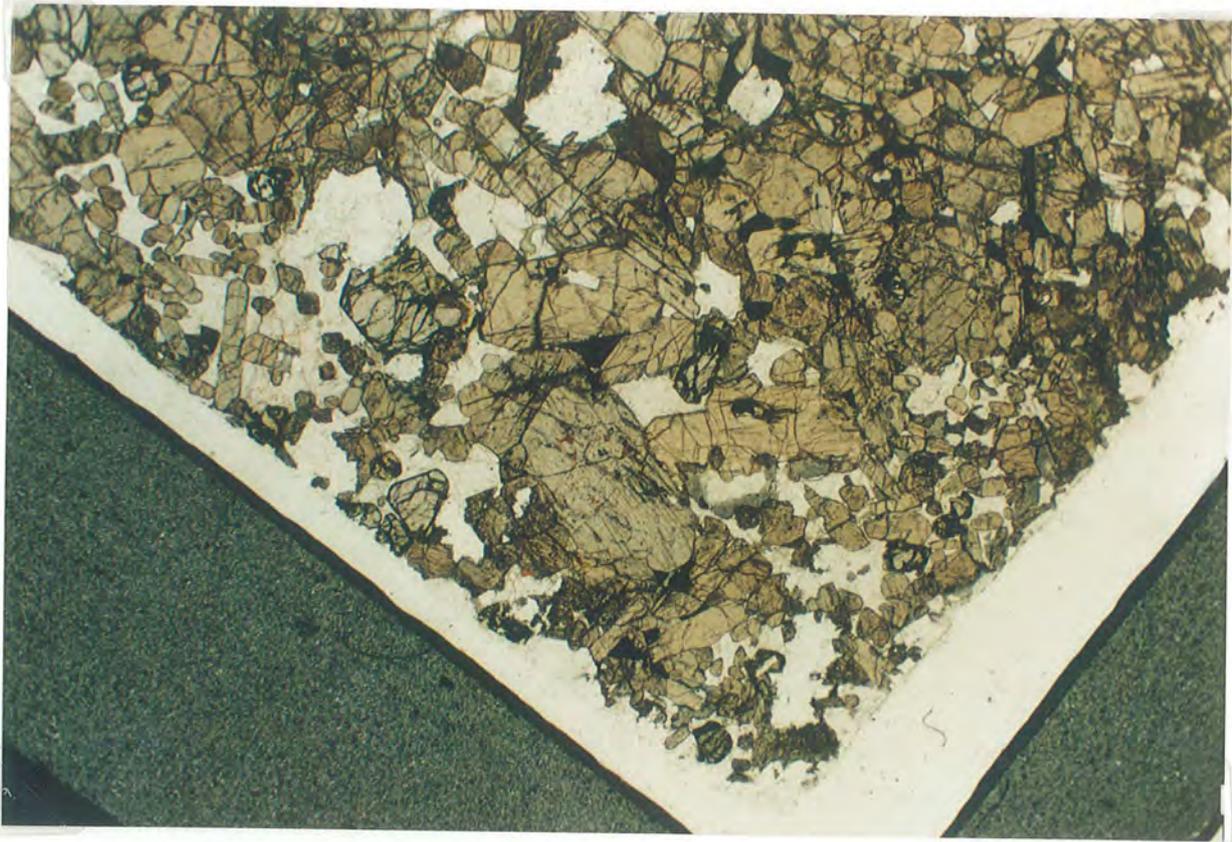


**Plate 3.100A.** Intensely corroded relict plagioclase porphyrocryst in thin pyroxenite band which replaces part of a porphyritic gabbronorite sheet ( along strike from section logged in Fig. 3.24A ). Sample 315D1. Crossed polarisers, width of field of view 3mm.

**Plate 3.100B.** Corroded relict plagioclase grain within poikilitic plagioclase. The latter encloses euhedral pyroxene crystals ( mainly orthopyroxenes ). From outcrop where the Upper Main Pyroxenite replaces porphyritic gabbronorites. Sample 6A. Crossed polarisers, width of field of view 1.5mm.



**Plate 3.101.** Upper Main Pyroxenite. General view showing variety of euhedral to subhedral pyroxenes ( mainly orthopyroxene ), rarer olivines and poikilitic feldspars. Rare high - grade metamorphic pargasitic hornblende is visible as inclusions in some large orthopyroxene grains. Plane - polarised light, width of field of view 14mm.



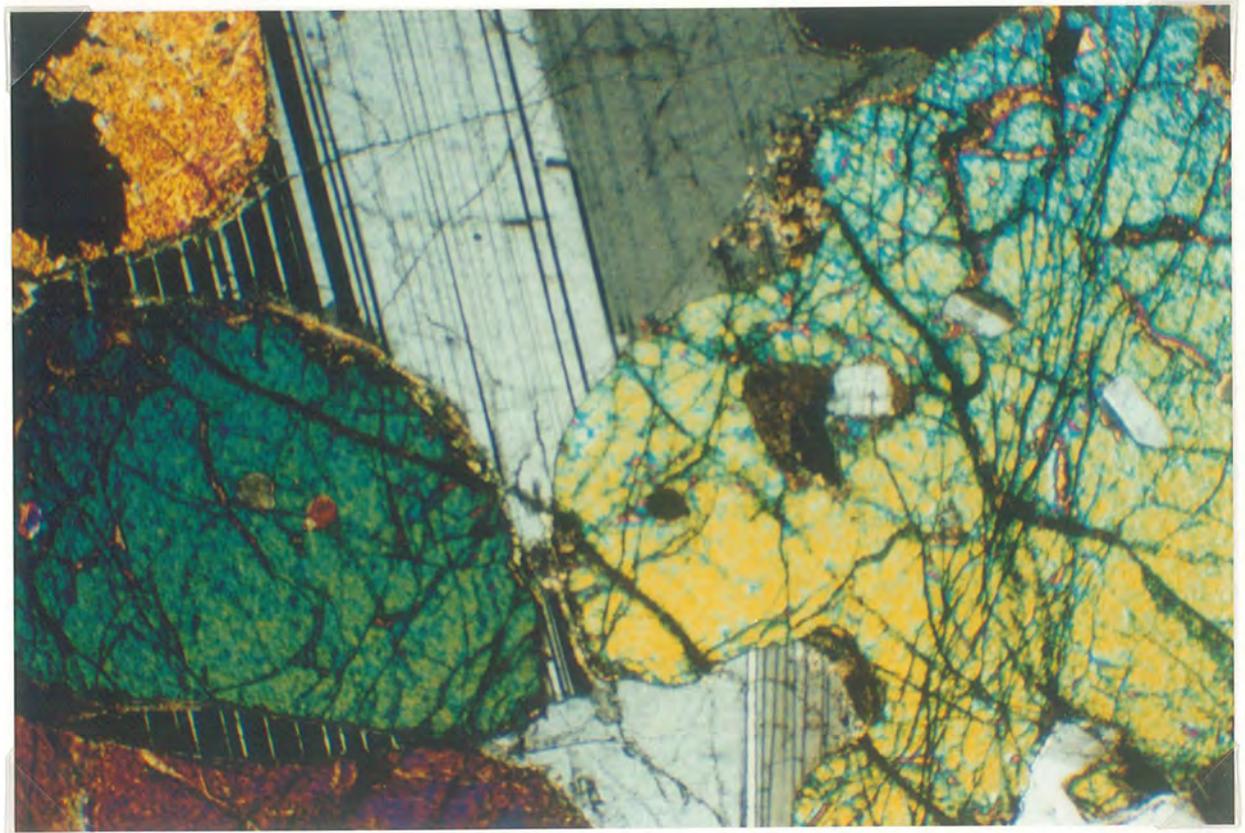
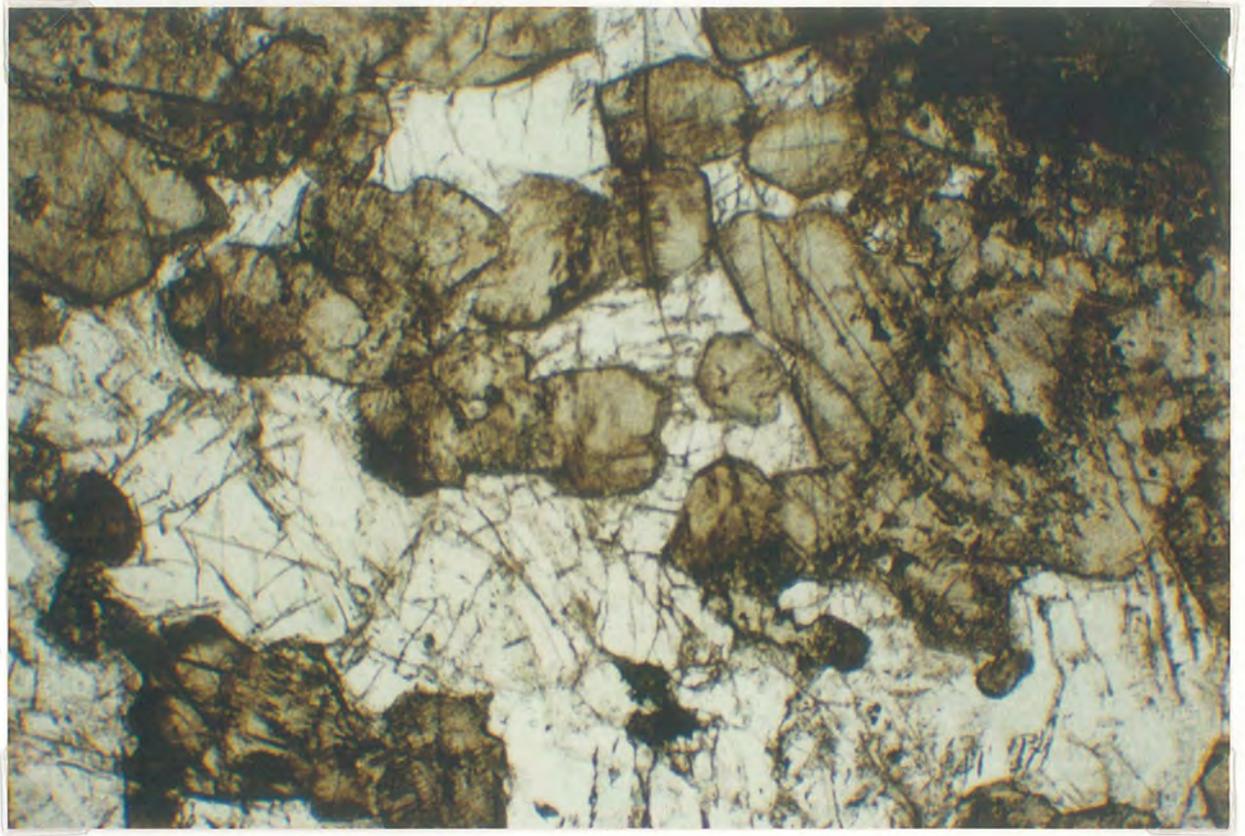
contact of both is a narrow transition zone of smaller ( 0.2 to 1mm ) euhedral to subhedral augites set in relatively abundant poikilitic plagioclase, which passes up into augite - rich isotropic to faintly laminated gabbros ( Plate 3.102 ). The one section of the lower contact ( from the LMP ) is markedly different, with a layer at least 2cm thick of embayed equant olivines enclosed in poikilitic feldspars at the top of the slide which pass down into interstitial laths as the contact is approached ( Plate 3.103 ). A notable feature of these rocks is the presence of textures, such as the subhedral pyroxenes and the poikilitic textures, which suggest an origin by crystallisation from a melt, rather than the granular or poikiloblastic ( in which the enclosed grains are typically rounded and anhedral ) textures which would indicate formation by solid - state or volatile fluid metasomatism. The exact nature of the metasomatic process is discussed further in Chapter 6. A final striking feature of these rocks, the much greater intensity of hydrous alteration, is described further in 3.4.6 and Chapter 6.

Overall, these replacive pyroxenites and related rocks bear a close resemblance to replacive or metasomatic rocks of various compositions which have been described from various intrusions. These include the Duke Island and Skaergaard complexes ( Irvine 1987 ), the Honningsvåg suite ( Robins et al. 1987 ), and possibly the Rhum complex as well ( Bedard et al 1988 ). The indistinct finger structures at the top of the pyroxene - rich gabbros are perhaps equivalent to much more strongly - defined finger structures in the Lille Kufjord ( Robins 1982 ), Rhum ( Butcher et al. 1985 ) and Honningsvåg ( Robins et al. 1987 ) complexes. However, the pyroxenites in Ardnamurchan differ from most of these occurrences in that they are not associated with interfaces between cumulates of differing compositions ( most usually between ultramafic and feldspathic cumulates, with replacement of the latter ) or between picritic sills and feldspathic cumulates, in the examples described from Rhum by Bedard et al. ( 1988 ). The Ardnamurchan pyroxenites, on the other hand, cross - cut earlier contacts and occur in a variety of rocks. In this they resemble the rocks described by Irvine ( 1987 ), but they have more regular ( tabular or planar ) geometries than the latter, which are completely irregular.

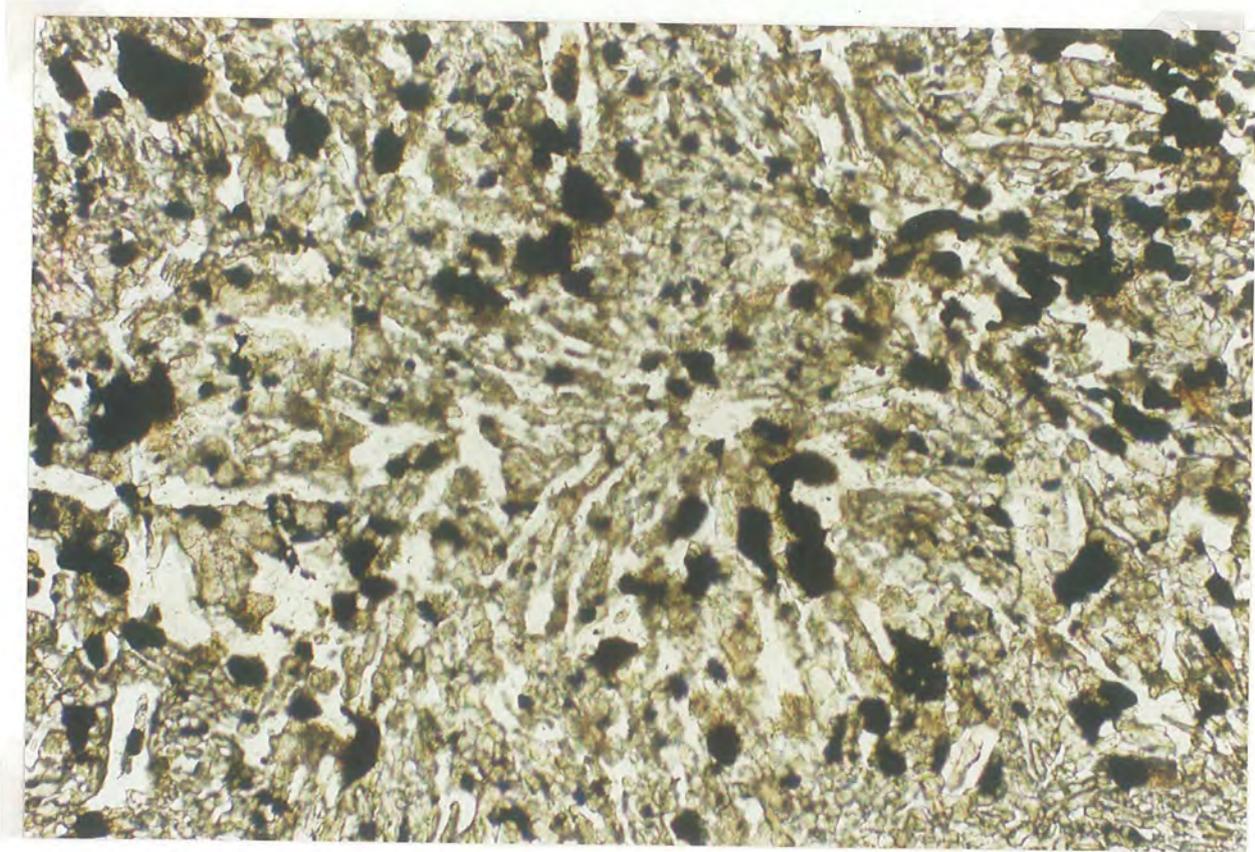
The age of secondary pyroxenite formation at Sanna is constrained to be younger than the emplacement of the gabbro-norite sheets there by cross - cutting relationships. The replacement of laminated rocks by the LMP and the overlying melagabbros ( Plate 3.99 and Fig. 3.28 ) indicates that the pyroxenite postdates the formation of lamination in the laminated gabbros. At 44057003, however, the LMP is deflected beneath an anorthositic xenolith which also deflects the lamination in the

**Plate 3.102.** Equant augite grains enclosed in poikilitic plagioclase at the top of the UMP ( within laminated gabbros ). Sample 316A. Plane - polarised light, width of field of view 3mm.

**Plate 3.103.** Embayed olivines enclosing small rounded plagioclase grains and themselves enclosed by poikilitic plagioclase. Base of LMP ( sample 313/3 ). Crossed polarisers, width of field of view 3mm.



**Plate 3.104.** Open spherulitic cluster of feldspar crystals, fine - grained hornfelsed granular dolerite. Sample 275/8, from vein cutting pre - gabbro norite laminated gabbros at An Acairseid. Crossed polarisers, width of field of view 1.5mm.



over- and underlying gabbros. Given the evidence that both the lamination ( see 3.4.2.1 ) and the pyroxenite are secondary in origin, this suggests that the pyroxenites formed towards the end of the period of formation of the lamination. This age relationship is consistent with the observation that some of the oikocrysts within the pyroxenites are strained or broken up into subgrains, suggesting slight deformation ( Plate 3.101, above ).

The Sanna area contains by far the largest concentration of secondary pyroxenites in the Inner Series. Other occurrences include uneven pyroxene - rich melagabbro bands in gabbronorites on Druim Reidh - Dhalach ( Grid Ref. 42936550 ) and xenoliths of augite - rich pyroxenite with poikilitic - interstitial feldspars in post - gabbronorite granoblastic dolerite veins at An Acairseid ( see section 3.4.4 ).

#### **3.4.4. Blastoporphyritic granular dolerite and microgabbronorite sheet intrusions.**

Fine - grained granular basic sheets, variously interpreted as xenolithic sheets or screens ( Richey et al. 1930, Wells 1954, 1978 ) or sheet intrusions ( Wells 1978 ), have been described previously from many parts of the Hypersthene gabbro. Whilst some of these can be shown to predate the adjacent gabbronorites ( section 3.4.2 ) others are clearly intrusive whilst still others have complex contact relationships with the gabbronorites. These later rocks are described in this section. Most are irregular but mainly inward - dipping sheets or tabular bodies, with most dips below 30°, even near the inner edge of the Inner Series, except on Beinn na Seilg ( dips of 40° to 70° ). A few radial dykes of similar coarse, unchilled granular dolerite outcrop at An Acairseid, with strikes of between 040° and 050°.

Planar sheets and less regular veins of granular dolerite, with sharp angular margins and containing angular xenoliths of their host rocks, occur in the laminated rocks at An Acairseid and cutting both gabbronorites and the Upper Main Pyroxenite at Sanna ( Grid Ref. 44077003 ). It appears from field relationships that these fine - grained intrusions were emplaced late on in the history of the Inner Series. This is consistent with their appearance in thin section. Although showing partial recrystallisation of the groundmass, relict igneous textures are present. Evidence for rapid crystallisation under moderate degrees of supercooling is present, in the form of elongate feldspar laths in sheaves and spherulitic clusters ( Plate 3.104 ). Chilled margins are absent, however,

and the pyroxenes and opaques in particular have recrystallised under pyroxene - hornfels facies conditions, hydrous minerals being absent. These intrusions are therefore beerbachites according to the definition of Phillips ( 1969 ). The presence of angular xenoliths and xenocrysts from the host rocks, and the sharpness and angularity of the margins of the intrusions, suggests that they were however emplaced under subsolidus conditions into brittle rocks.

Higher host rock temperature, and hence more fluid rheology, may account for the complex contacts of a similar but coarser - grained group of basic sheet intrusions. These have irregular, intensely veined unchilled margins and are found throughout the Inner Series, but are especially common in swarms on Beinn na Seilg and on Stacan Dubha, around 464642. The marginal veins are formed by isotropic gabbroic rocks which are usually indistinguishable in the field from the host gabbroites ( Plate 3.105 ). Where they occur in laminated gabbros, however, they are enclosed in an envelope of isotropic gabbro which grades outward into laminated gabbro and passes inwards into the veins which penetrate the unchilled intrusions ( Plate 3.106 ). This indicates that the veins are in fact backveins, produced by remelting and mobilisation of the host rocks by the intruding magma. The later age of the sheets is confirmed by the presence in them of host - rock xenoliths. The occurrence of back - veining in these rocks suggests that the intruded sheets had reached a sufficiently advanced stage of crystallisation to become a congealed mass of crystals, with interstitial melt, at a temperature below which the immediately adjacent host rocks formed a layer of mobile magma a few centimetres thick.

Although the field evidence indicates that they were intruded into rocks at higher temperatures than the beerbachitic dolerites, these microgabbroites have relatively well - preserved ophitic textures, although the olivines in particular occur as clusters of subgrains produced by disruption of interstitial oikocrysts, suggesting that some textural reequilibration occurred in these rocks. The discrepancy between temperature of host rocks and degree of recrystallisation of the two groups of intrusions can be explained in terms of the grain size dependence of recrystallisation rates ( Vernon 1970 ).

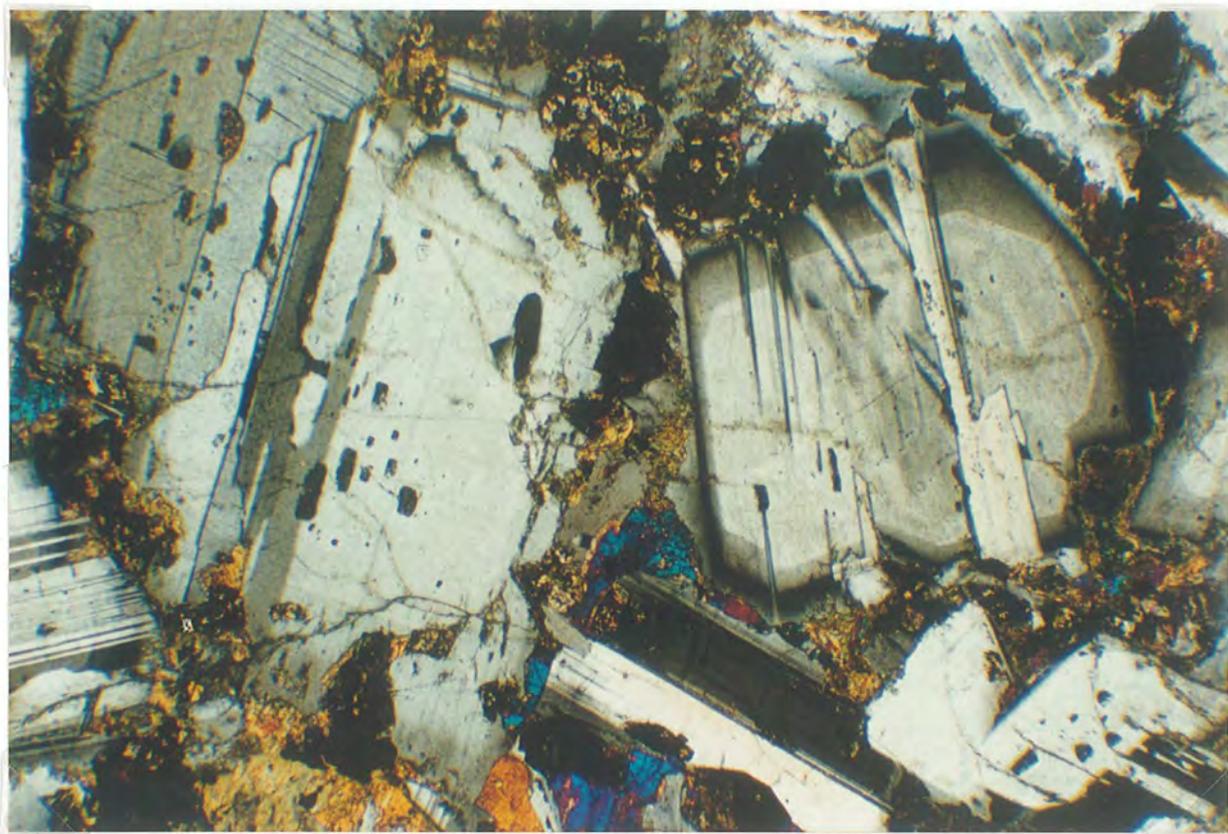
The mineral assemblage in both the microgabbroites and in the beerbachites is tholeiitic, although orthopyroxene is much less abundant in the latter. Olivine is also invariably present, although in varying amounts, but is usually partly replaced by opaques or opaque - pyroxene intergrowths: these

**Plate 3.105.** Lobate margin of microgabbroite sheet emplaced into coarser isotropic gabbroites, Beinn na Seilg. Field location 279/5; grid reference 464642 approx.

**Plate 3.106.** Discontinuous microgabbroic sheet with lobate margins intruded into laminated gabbros, Sanna ( field location 311, grid reference 44007004 ). The host rocks appear to lose their lamination within a few centimetres of the sheet, suggesting that they were partially melted by it.



Plate 3.107. Plagioclase porphyrocrysts with granular opaque - augite inclusions, from Group A feldspar - phyric dolerite ( sample 308 ). Crossed polarisers, width of field of view 3mm.



are the youngest rocks to be affected by the pervasive high - grade metamorphism that occurs in the Inner Series ( see section 3.4.6 ). Both rock types are plagioclase - phyric. The porphyrocrysts are of two types, elongate euhedral zoned phenocrysts and irregular, internally strained xenocrysts, probably restitic in origin. Polycrystalline feldspar - rich microxenoliths are also present.

### 3.4.5 Post - metamorphic minor intrusions.

All the Inner Series rocks described in the preceding sections predate, or were emplaced during, the high grade metamorphism discussed in section 3.4.6, below. A number of minor intrusions were emplaced into the Hypersthene Gabbro after this metamorphism, some of which were identified as such by Richey et al. ( 1930 ). During the course of the present work these minor intrusions were found to be more common and varied than had previously been recognised. They fall into three main groups:

1): **Group A).** Coarse, highly feldspar - phyric dolerites. These are as much as 60 - 70% subidiomorphic, near - equant plagioclase crystals and coarse anorthositic microxenoliths. These feldspars all have corroded, unzoned cores with strongly zoned rim overgrowths and rounded inclusions of fine - grained granular augite - opaque aggregates, which may represent recrystallised melt pockets ( Plate 3.107 ). Rare large, equant augite grains with interstitial overgrowths are also present but the bulk of the rest of these rocks is an interstitial fine - grained groundmass of plagioclase laths, ophitic to granular augite, accessory opaques and either rare olivine or equally rare quartz - alkali feldspar intergrowths. Orthopyroxene is absent. Intrusions of this type outcrop around 423669 ( a shallowly east- ( inward- ) dipping sheet some 5m thick ), in a fault - bounded outcrop around 454639, and in several shallowly north - dipping sheet intrusions on Beinn na Seilg. In the latter two areas the ultraporphyritic dolerites are cross - cut by granophyre veins and mixed - magma intrusions of group C, below.

2): **Group B.** Porphyritic fine - grained dolerites, with strongly chilled margins and, in some, strongly feldspar - phyric axial zones. These are petrographically indistinguishable from the porphyritic dolerites found in the MBG and outside the pluton which belong to the Group 4 intrusions defined in section 3.2.5.4. They form part of a swarm running across the southern part of the pluton and the adjacent country rocks, composed of steeply north - dipping sheets and subvertical dykes,

with strikes of 070° to 100°. On Beinn na Seilg, some of these pass along - strike into hybrid sheets of group C, whilst others cut the Beinn na Seilg quartz gabbro ( as defined by Richey et al. 1930 ) to the north, around 45826423.

**3): Group C.** A large number of granophyre and hybrid granophyre / dolerite pillowed and net - veined sheets, typically either sub - horizontal or sub - vertical occur in many parts of the Inner Series. They are usually associated with felsic joint - like veins in the adjacent rocks. Those outcropping close to the inner margin of the pluton as it is now exposed can be traced laterally into the Sgurr nam Meann hybrid dolerite / granophyre intrusion ( described in detail in Marshall ( 1984 )), particularly on Beinn na Seilg.

These various post - metamorphic minor intrusions all lack orthopyroxene and do not seem to be geochemically related to the mainly strongly tholeiitic Inner Series rocks. Groups B and C seem to be related to the Sgurr nam Meann intrusion, although rigorous investigation of this hypothesis would require an extensive programme of analysis of the various minor intrusions outcropping away from the inner edge of the Hypersthene Gabbro which is beyond the scope of the present work. The earlier group A has similar age relationships to those of the post - metamorphic feldspar - rich gabbros in the Glebe Hill area and are petrographically similar: this possible relationship is examined further in Chapter 6.

#### **3.4.6 Metamorphism of the Inner Series rocks.**

The boundary between late magmatic and subsolidus processes in many mafic plutonic rocks is blurred by the fact that a major texture - forming process, textural reequilibration, occurs in both partially molten ( Hunter 1987 ) and high - grade metamorphic rocks ( Kretz 1966; Vernon 1970 ). However, the rate of textural equilibration, ( recrystallisation driven by the release of excess surface energy associated with lattice mismatches at grain boundaries ) is governed largely by the rate at which material can be dissolved, moved along grain boundaries, and re - precipitated elsewhere. Rates of these processes, and hence the rate of textural equilibration, are poorly constrained but are probably much higher when melt is present than in subsolidus rocks, particularly when the rocks concerned are polymineralic ( Hunter 1987 ). Thus for any given time - scale available for the operation of textural re - equilibration, the maximum length - scale over which re - equilibration can

occur, and hence the largest grain size of primary, non - equilibrium textures which can be affected by it, will be much less in rocks just below the solidus than in those just above. This reduced rate of material transport will also be apparent in the much finer grain size of the original textures. Consequently, even if the minerals involved are the same, it is in theory possible to distinguish late magmatic from subsolidus textures simply by the much finer grain size and reduced degree of textural equilibration typical of the latter, unless a particularly reactive non - silicate fluid phase is present.

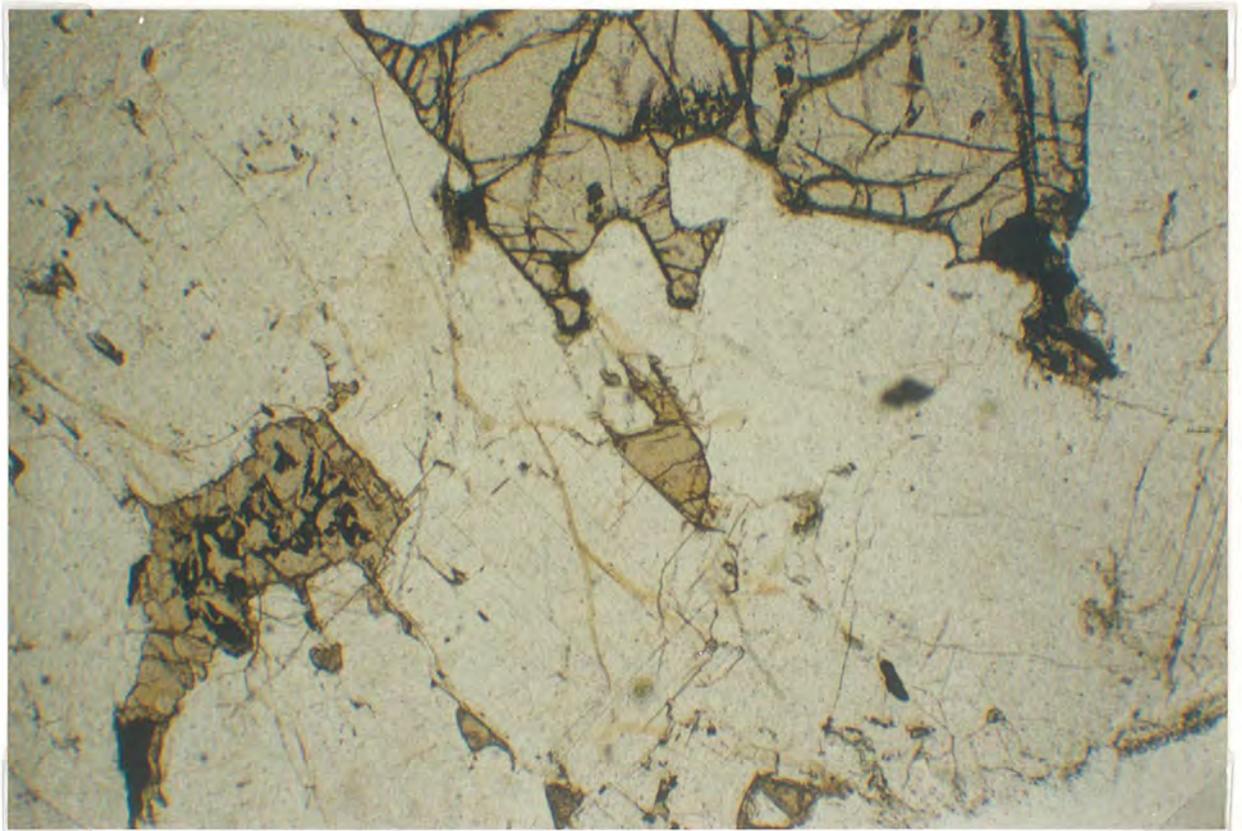
Evidence of this type for a distinct phase of very high temperature sub - solidus metamorphism in the Inner Series of the Hypersthene Gabbro is presented in this section. Detailed mineralogical evidence to support the subsolidus origin of these textures and the anhydrous mineral assemblages involved is presented in Chapter 6, as are quantitative estimates of the conditions of metamorphism.

The textures and microstructures considered to form evidence of subsolidus metamorphism are of three types: replacement textures ( particularly those produced by partial oxidation of olivine ), exsolution - recrystallisation textures, and joint - like veins and the alteration haloes which occur around them.

An example of the most striking replacement texture is shown in Plate 3.108. This is a symplectitic intergrowth of orthopyroxene and opaques which normally occurs as a patchy partial replacement of olivine. Where the olivine is rimmed by interstitial magmatic orthopyroxene, the orthopyroxene in the intergrowth is in optical continuity with the latter. Patches of opx - opaque intergrowth sometimes occur without relict olivine, in the cores of orthopyroxene grains. These may have been produced by complete replacement of olivine grains, at least within the plane of section, but similar symplectites have been described from olivine - free rocks ( Barton & Van Gaans, 1988 ). In the Inner Series, however, those symplectitic intergrowths not associated with relict olivine consistently occur within orthopyroxene grains. This suggests that they may be the product of complete replacement of olivine, which is usually rimmed by interstitial opx of apparently ( because of its coarse texture ) late magmatic origin. The opaque phase in these intergrowths is usually titanomagnetite, although a few ilmenite - orthopyroxene intergrowths were found ( see section 6.4 ). These symplectites were found in about 80% of 30 olivine - bearing Inner Series rocks ( mainly gabbronorites but also laminated gabbros, anorthosites and late microgabbronorite sheets ) examined in detail in thin section. The

**Plate 3.108.** Orthopyroxene - opaque oxide symplectitic intergrowth replacing olivine in gabbro-norite ( sample 35 ). A lamellar structure is present in parts of this intergrowth but they are usually less regular. Crossed polarisers, width of field of view 1.5mm.

**Plate 3.109.** Coarse - grained granular orthopyroxene - opaque oxide intergrowth ( at left - hand side of plate ). This is interpreted as having formed by recrystallisation of a symplectitic intergrowth which had replaced part of the olivine poikilocryst at the right - hand side of the plate. Sample 287C1. Plane - polarised light, width of field of view 3mm.



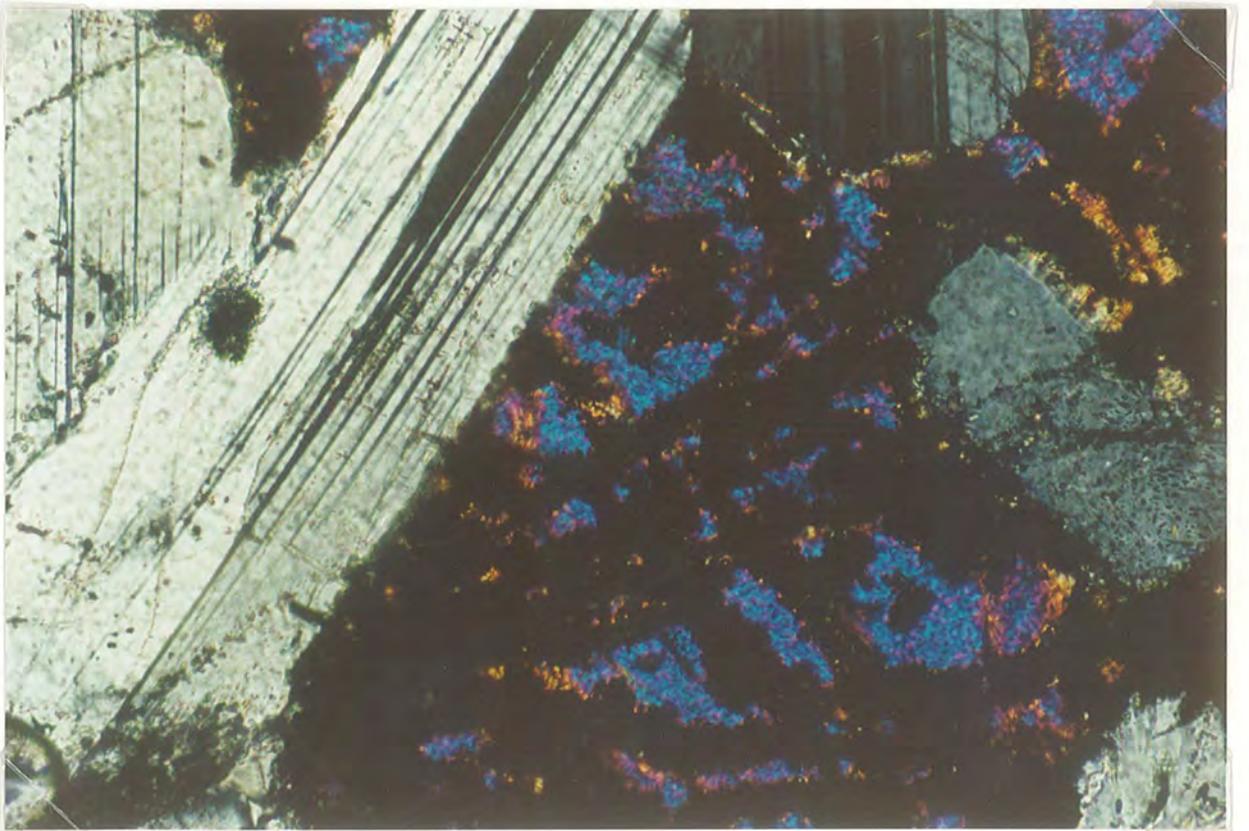
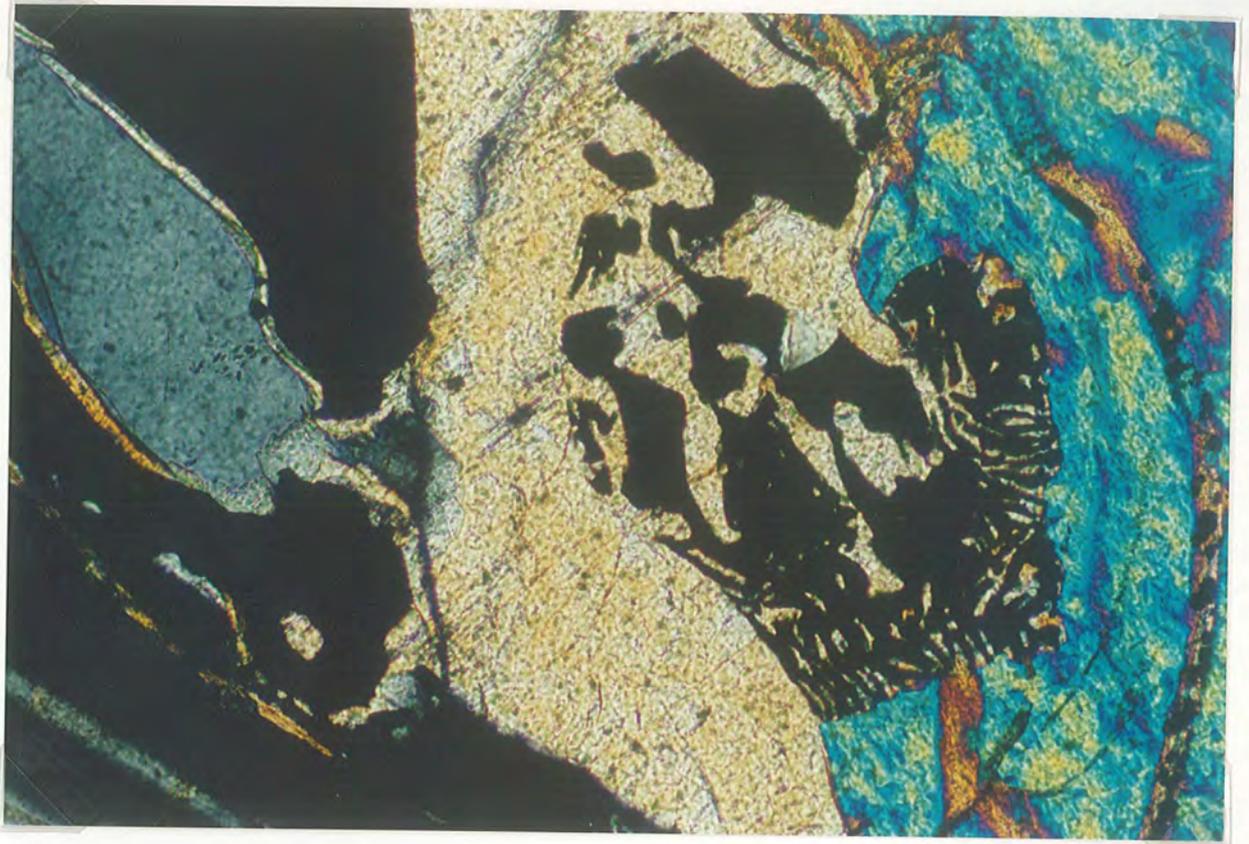
proportion of symplectitic intergrowth present in any one rock is, however, small: at most 2 - 3% by volume, and usually much less, even in rocks with 10 - 15% primary modal olivine. Although the symplectite texture is often well - preserved, coarser irregular intergrowths of orthopyroxene and opaques are also present ( Plate 3.109 ). In a few examples the symplectitic intergrowth shows a fine - scale intergrowth close to the interface with the olivine which it replaces, whilst away from the contact the intergrowth is coarser and less regular, suggesting that the older material has undergone grain coarsening through recrystallisation ( Plate 3.110 ). By analogy, the coarser orthopyroxene - opaque intergrowths elsewhere may also have formed as symplectitic intergrowths which have subsequently undergone partial textural equilibration.

Another olivine oxidation - replacement texture is found mainly in rocks with high contents of coarse poikilitic and texturally equilibrated equant opaque grains ( in other words, in those rocks with a high content of opaques prior to olivine replacement ). In these, the olivine is partly or wholly replaced by very fine granular oxides. These may occur disseminated throughout the olivine grains or, more commonly, are concentrated along grain boundaries and fractures in the olivines ( Plate 3.111 ). Very rarely, some of the olivine in these rocks is replaced by orthopyroxene choked with opaque inclusions. Augites in these rocks also contain a few opaque granules but are not affected to anything like the same degree. Other, less oxide - rich rocks contain olivines with granular oxides in their cores and discontinuous, granular oxide - orthopyroxene rims. The replacement of olivine appears to have been unreversed in almost all of the Inner Series rocks examined. However, a reduction reaction which may correspond to the olivine - oxidation reaction was found in one magnetite - anorthosite rock, sample 166B2 from the banded xenolithic sheet south - west of Beinn na Seilg ( section 3.4.2.3, above ). In this rock, some coarse - grained magnetite grains dating from the formation of the granular texture are fringed by a myrmekite - like intergrowth of plagioclase and a ferromagnesian mineral ( olivine or clinopyroxene ). The formation of this texture is discussed further in section 6.4.2.2.

Although two pyroxenes are present in most IS rocks, classical exsolution textures are not always developed except in inverted pigeonites. This is particularly the case in augites, which sometimes contain orientated needles and lamellae of opaques in place of orthopyroxene. Patchy recrystallisation of these to coarser equant granules of opaques in clear augite is common ( Plate 3.112 ). This patchy recrystallisation is particularly intense close to veins and annealed fractures ( see below ).

**Plate 3.110.** Small area of symplectitic intergrowth replacing olivine ( sample 137/1 ). The intergrowth shows progressive coarsening away from the reaction front. Crossed polarisers, width of field of view 0.75mm.

**Plate 3.111.** Olivine replaced by granular opaque oxides alone. Laminated granular magnetite - anorthosite, sample 284. Crossed polarisers, width of field of view 0.75mm.



The reactions involved in the formation of these textures and the conditions of their formation are discussed in detail in section 6.4. However, the abundance of  $\text{Fe}^{3+}$  - bearing oxide phases amongst the products points to the occurrence of oxidation reactions during metamorphism: this, and the evidence to be presented in Chapter 6 for open - system behaviour of other elements besides oxygen, indicates that these replacement and oxidation reactions took place in the presence of a fluid phase. Given the argument above to the effect that if this fluid were a silicate melt, large - scale textural re - equilibration would probably have destroyed the delicate reaction textures, this is more likely to have been a very high - temperature aqueous hydrothermal fluid. The same conclusion also results from the mineralogical studies and mineral geothermometry dealt with in Chapter 6.

More petrographic evidence for the involvement of aqueous fluids in the high - temperature metamorphism of the IS is present in the form of a dense network of near - parallel - sided joint - like veins containing anhydrous mineralogies. These veins range in thickness from 1 - 1.5mm to as little as a few microns: most are less than 0.2mm thick. The spacing of the veins varies with size from as much as a few tens of centimetres ( even more in the case of the rare mm - sized veins ) down to less than a centimetre. About one - third of normal - size thin sections ( 3cm by 2cm ) contain one or more veins in the 0.2 - 0.05mm thickness range, and the thinnest, oxide - rich veins occur in large numbers in almost all Inner Series rocks.

The veins contain a variety of assemblages: orthopyroxene + opaques, with or without small amounts of augite and / or plagioclase; orthopyroxene only; augite + minor plagioclase and opaques; opaques only. Opaque - rich assemblages are especially common in the smallest fractures. Biotite, green hornblende and other hydrous phases are often present, but only as later replacements of the anhydrous phases or as distinct zones in veins with a multistage history ( Plate 3.114, below ). Although the earlier vein assemblages are anhydrous, this does not exclude the possibility of their formation from, or by reaction with,  $\text{H}_2\text{O}$  - rich fluids at high temperatures and low  $P_{\text{H}_2\text{O}}$  ( Spear 1981; see Chapter 6 ). Several textural features of the veins indicate that they did not crystallize from silicate melts: they do not contain cotectic or near - cotectic proportions of their component phases; they do not show textural evidence of either rapid crystallisation or permeation into the host rocks, which would be expected to occur if they had crystallised slowly ( Hunter 1987 ); most of the veins show little sign of textural equilibration or annealing except on scales of less than about 10 microns. As is the case with the symplectites and other replacement textures, however, the best evidence for the

hydrothermal origin of these veins comes from detailed mineralogical studies of their conditions of formation.

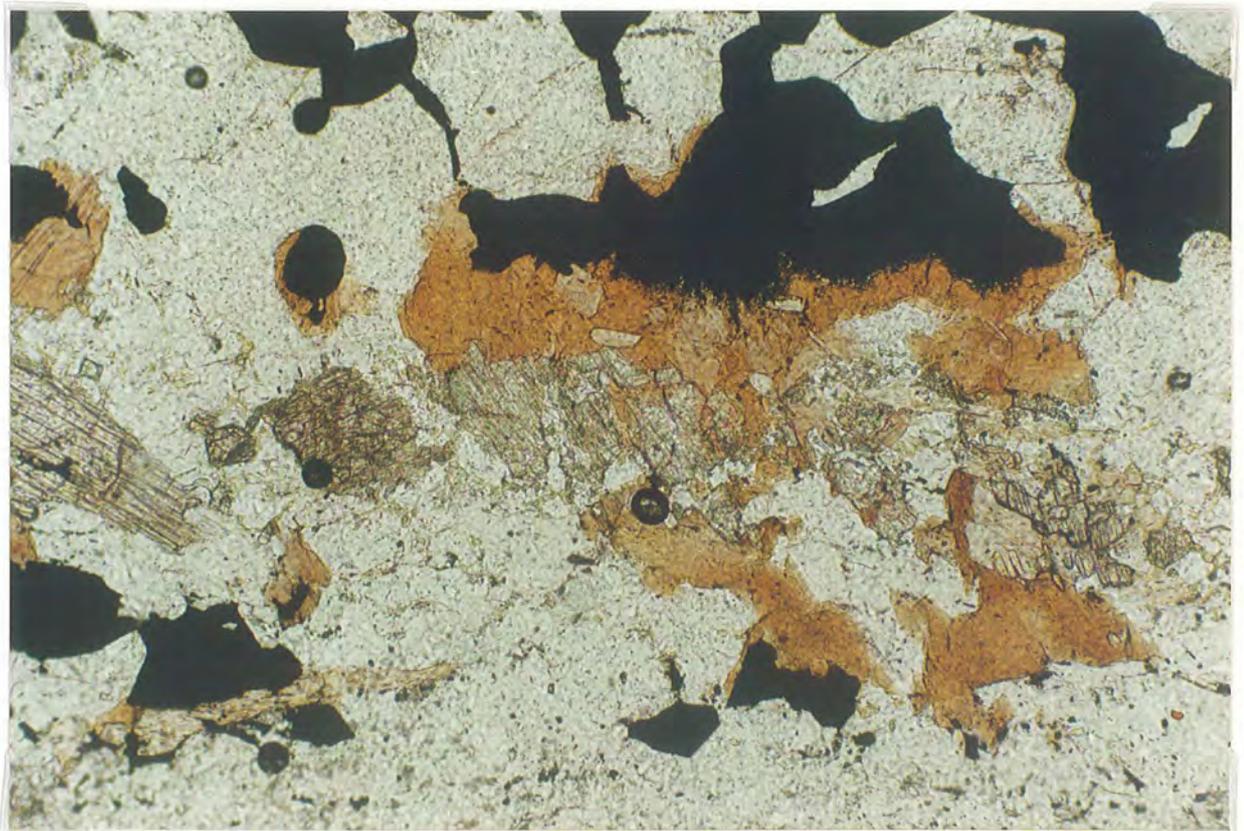
The vast majority of these fractures show no evidence of deformation by shearing across them and no offset of earlier structures cut by them ( for example, a narrow olivine - rich band running vertically across Plate 3.113 shows no offset across the orthopyroxene - rich vein which cuts it ). Those fractures which are associated with shearing displacements are highly distinctive ( see section 3.4.7 ) and are found only near major faults. **This suggests that almost all of the high - temperature fractures which cut the Inner Series of the Hypersthene Gabbro form a tensile fracture network similar to that seen in the contact aureole and outer margin of the Marginal Border Group, but which formed at a much higher temperature. The small - scale partial annealing and textural equilibration seen in some of the fractures, and the high - temperature mineral assemblages which plug them, suggest that fracture formation in these rocks was not followed by a rapid drop in temperature, unlike the situation in the MBG and the surrounding rocks.**

Fracture formation seems to have persisted at lower temperatures within the Inner Series, as later fractures which contain only hydrous assemblages ( biotite  $\pm$  opaques; biotite + green hornblende  $\pm$  opaques  $\pm$  haematite; hornblende + chlorite  $\pm$  haematite, chlorite  $\pm$  epidote  $\pm$  sericite ) occur in addition to extensive overprinting of the anhydrous veins by the same assemblages. A common feature of the larger hydrous veins is a zonation from biotite - rich margins to hornblende - chlorite - haematite cores which may represent a temporal variation due to falling temperature with time in a long - lived fracture ( Plate 3.114 ). A notable absentee from the vein assemblages is brown pargasitic hornblende, which does however occur in the host rock ( see below ), suggesting slight variations in the conditions of high grade metamorphism which are discussed in Chapter 6.

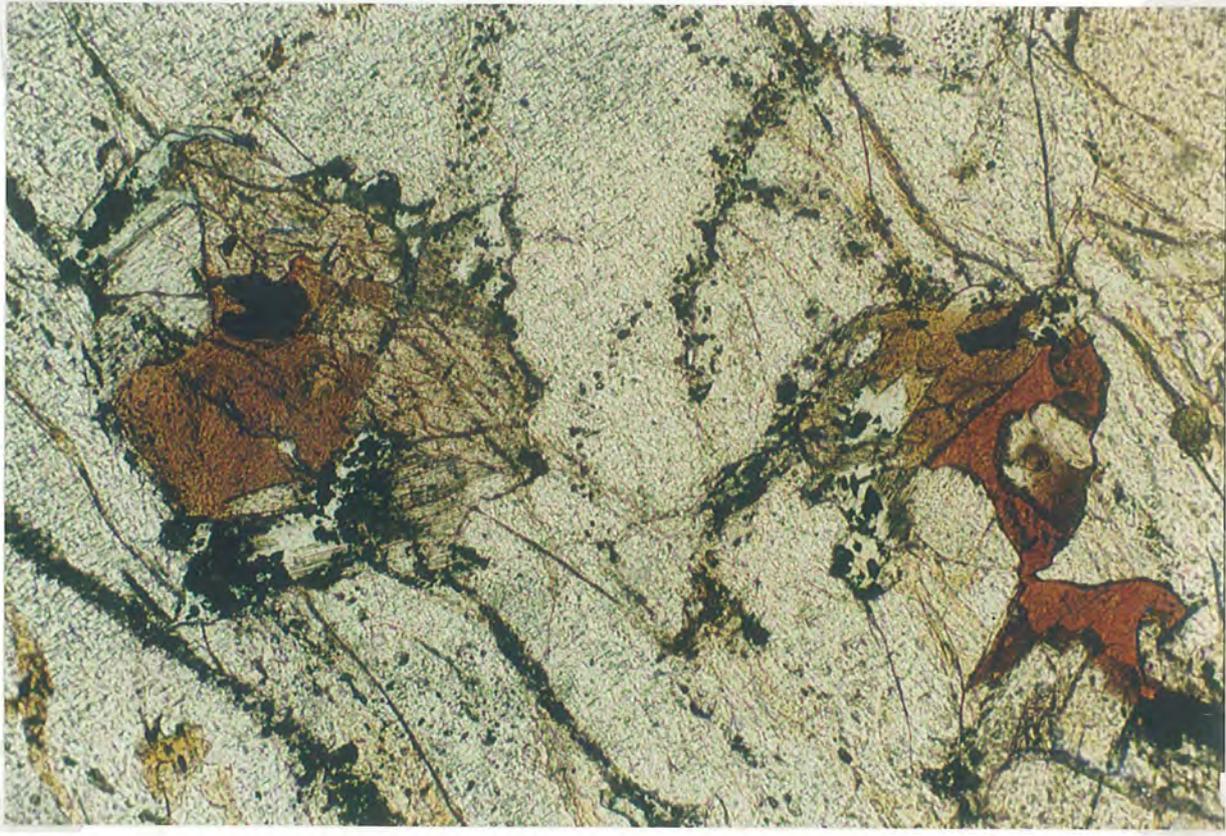
Overprinting of anhydrous, magmatic and high - temperature metamorphic mineral assemblages by lower grade hydrous assemblages is ubiquitous in the Inner Series but is usually very incomplete ( even the most unstable phase, olivine, is largely preserved in most sections ), although the degree of development varies widely. The following are the main types of alteration, in a sequence which is, broadly speaking, one of decreasing age in these rocks as well as one of decreasing metamorphic grade:

**Plate 3.113.** Annealed orthopyroxene - rich vein in sample 97/1 ( see Plate 3.91 ). An olivine - rich band running across the centre of the plate is not offset across the vein, suggesting that it initially formed as a tensile fracture. Plane - polarised light, width of field of view 14mm.

**Plate 3.114.** Zoned hydrothermal vein in magnetite - anorthosite ( sample 166A2 ) with relict orthopyroxene in the amphibole - bearing core of the vein and biotite - rich margins. Plane - polarised light, width of field of view 1.5mm.



**Plate 3.115.** Composite inclusions from within a large olivine grain, sample 137/2. The inclusions contain orthopyroxene + pargasitic hornblende  $\pm$  Ti - biotite  $\pm$  magnetite. The host olivine is also cut by annealed fractures decorated with opaque grains. Plane - polarised light, width of field of view 1.5mm.



1). Formation of titaniferous biotite rims on primary opaques and of pargasitic hornblende grains. The latter are especially abundant in the anorthosites and secondary pyroxenites, and almost invariably occurs only as inclusions in larger anhydrous mineral grains, for reasons to be discussed in Chapter 6. In some cases, the hornblende may coexist with one or more pyroxenes ( Plate 3.115 ). These minerals are usually only present in trace amounts.

2). Replacement of olivine by talc and granular opaques. This is pervasive in some rocks but is normally patchy and incomplete, and particularly intense close to biotite - hornblende bearing fractures.

3). Replacement of a few percent of orthopyroxene grains by actinolite or chlorite, and even less common replacement of augite by chlorite  $\pm$  green hornblende. Rare patchy replacement of calcic feldspar by epidote.

4). Low - grade minerals, usually only developed along rare late fractures: smectite, serpentine and goethite ( after olivine ), carbonate and zeolites.

On the outcrop and larger scales the distribution of alteration is rather uneven. The rock types most prone to alteration are the secondary pyroxenites, which show extensive development of chlorite and green hornblende, and prominent but rare pargasitic hornblende and biotite, but little replacement of olivine by talc, and the coarse anorthosites, in which pargasitic hornblende and later, much lower - grade epidote and zeolites are more common than usual, although still forming only 1 - 2% of the rock as a whole. On the map scale, hydrous alteration in the IS is most intense in the vicinity of the major NNW - SSE trending fault zones in the west of the peninsula, the radial faults at Briaghlann, An Acairseid and on either side of Beinn na Seilg, and towards the inner and outer margins of the Inner Series as it is presently exposed. The concentration of alteration around the fault zones suggests some degree of concentration of fluid flow along them on a large scale, although as noted above the tensile fractures appear to have been more important as fluid flow conduits on the thin section scale.

The development of particularly intense hydrous metamorphism towards the inner edge of the Inner Series is to be expected because this contact is formed by later intrusions. The effects of their

hydrothermal systems will therefore be superimposed on that due to self - metamorphism by the cooling Hypersthene Gabbro. It should be noted, however, that the age relationships of the late - metamorphic NNW - SSE trending faults cutting the Inner Series indicate that the majority of the metamorphism, and all of the high - temperature metamorphism, described here must predate these intrusions because the faults also do so ( see section 3.4.7 ).

The faults which form much of the outer contact of the Inner Series juxtapose IS rocks with well - developed high - grade, pyroxene - hornfels facies metamorphism of the type described above, against MBG rocks affected by M2 metamorphism of markedly lower grade ( with replacement of olivine by talc and orthopyroxene by actinolite or hornblende ). However, where intrusive contacts occur between Inner Series and MBG rocks, for example on the western side of Beinn nan Codhan around 438635, on the eastern and western sides of An Acairseid ( again, see Map 4 ), and at the southern end of Gharblach Mhor, the adjacent Inner Series rocks show a rather atypical pattern of metamorphism. Orthopyroxene - opaque symplectites are almost completely absent from these rocks and high - temperature anhydrous vein assemblages were not found. Recrystallisation of magmatic augite to clear augite with opaque inclusions does occur in these rocks, as it does in the inner parts of the M2 aureole outside the Inner Series, but experimental constraints and observations from other intrusions ( see Chapter 6 ) suggest that this reaction will occur at lower temperatures than other anhydrous reactions. The remainder of the metamorphic mineral assemblage in these IS rocks includes unusually well - developed replacement of olivine by talc + opaques, titaniferous biotite rimming and replacing primary opaques, and patches of brown pargasitic hornblende. This more moderate grade metamorphism, at high hornblende - hornfels facies, is comparable to the highest grades of M2 metamorphism in the adjacent MBG and country rocks. The continuity in conditions of metamorphism suggests that emplacement of these Inner Series rocks may have taken place during M2 in the surrounding rocks, which raises the possibility that part of M2 in the MBG and country rocks was caused by the Inner Series. Emplacement of the Inner Series into rocks already at moderately high temperatures is however necessary, to explain the lack of strongly chilled margins at the intrusive contacts between the Inner Series and the MBG.

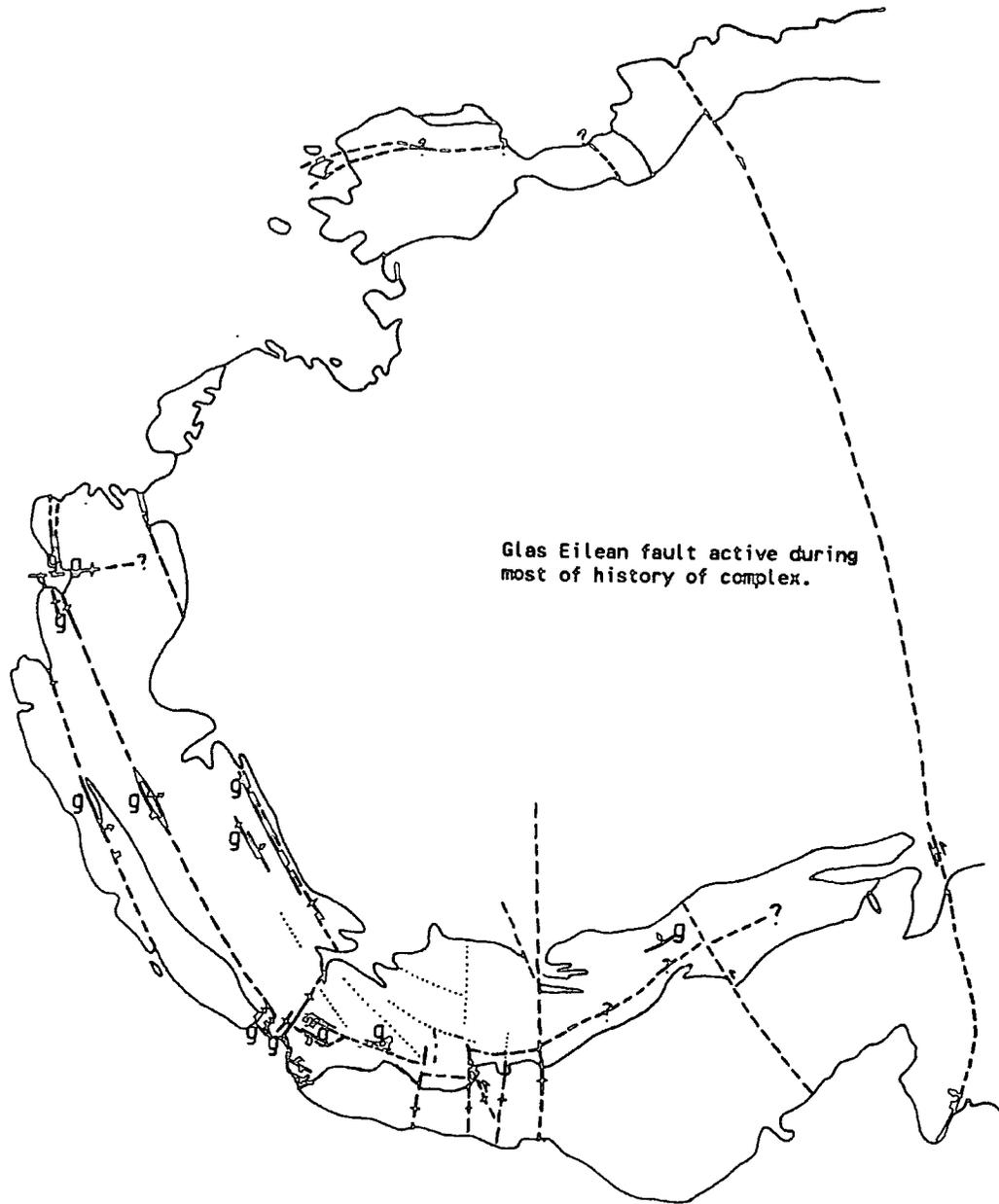
#### **3.4.7. Deformation of the Inner Series and criticisms of the central subsidence model for layering orientation.**

Two types of deformation can be distinguished within the Inner Series. The earlier of these is the pervasive, late magmatic deformation seen in the laminated gabbros and the gabbronorites, which to a large extent is expressed as the development of laminated fabrics. This has already been dealt with in sections 3.4.1 and 3.4.2. The second type of deformation is almost entirely confined to discrete, mappable fault zones. These include the concentric faults which form the outer margins of the Inner Series in many places, but also a number of faults within it, which are also shown on Fig. 3.29.

Although of various overall geometries, all these faults, where exposed, are much the same in being composed of broad zones, up to 100m wide, of fine- to medium- grained granular basic rocks with a weak, steeply - dipping fabric defined by wispy bands of finer- and coarser- grained rock a few millimetres wide. In thin section, the finer - grained rocks are granoblastic, polygonal - textured microgabbronitic to microtroctolitic rocks with a well - defined preferred crystallographic orientation with plagioclase z - axes parallel to the fault zones, but little or no morphological preferred elongation of crystals ( Plate 3.116A and B ). This suggests that intense static recrystallisation of an initially much finer - grained fault rock has taken place. The coarser - grained rocks sometimes retain a preferred elongation of feldspars in the plane of the banding and the fault itself, despite also being extensively annealed. Some of the larger grains show intracrystalline straining, suggesting the operation of intracrystalline deformation, although the very well - developed preferred orientation in the finer - grained rocks points to deformation by recrystallisation processes, with preferential growth of crystals in the plane of the fault, after initial intense grain - size reduction. As with the texturally very similar shear zones in the hanging walls of the faults bounding the Inner Series, the preservation of anhydrous mineral assemblages ( particularly olivine and two pyroxenes ) points to deformation at very high temperatures indeed, above the amphibole stability field ( see Chapter 6 for quantitative estimates ). By analogy with the faults at An Acairseid ( section 3.3.2 ) this deformation probably took place during the latter part of the period of emplacement of the gabbronorite sheets, although because of the intermittent exposure and the difficulty of identifying internal contacts within many of the gabbronorites no actual cut - offs were observed.

Later movements, at lower temperatures, also occurred on most of these fault zones. This later strain appears to have been concentrated on progressively narrower faults or fault sets, as later deformation and hydrous alteration at lower temperatures is concentrated on discrete zones which usually erode preferentially to form striking linear gullies in which isolated outcrops of intensely deformed rocks

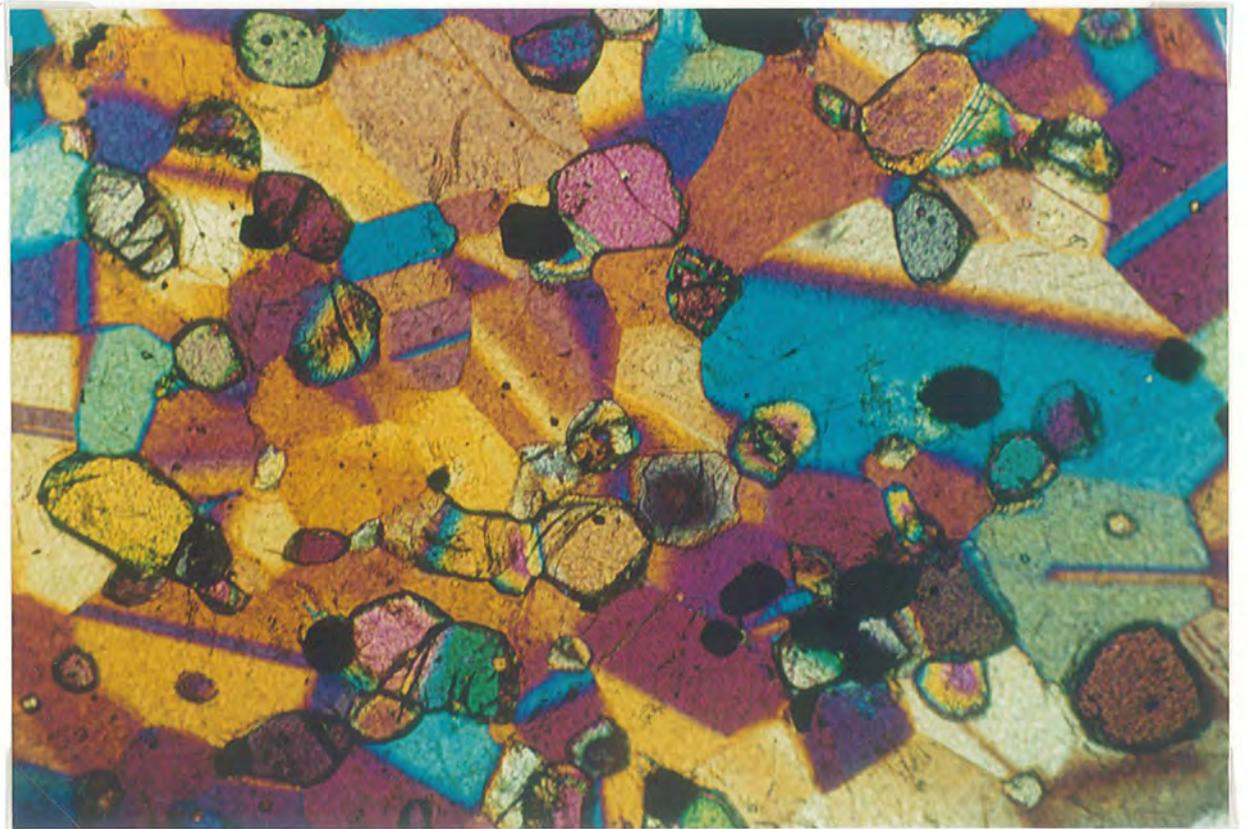
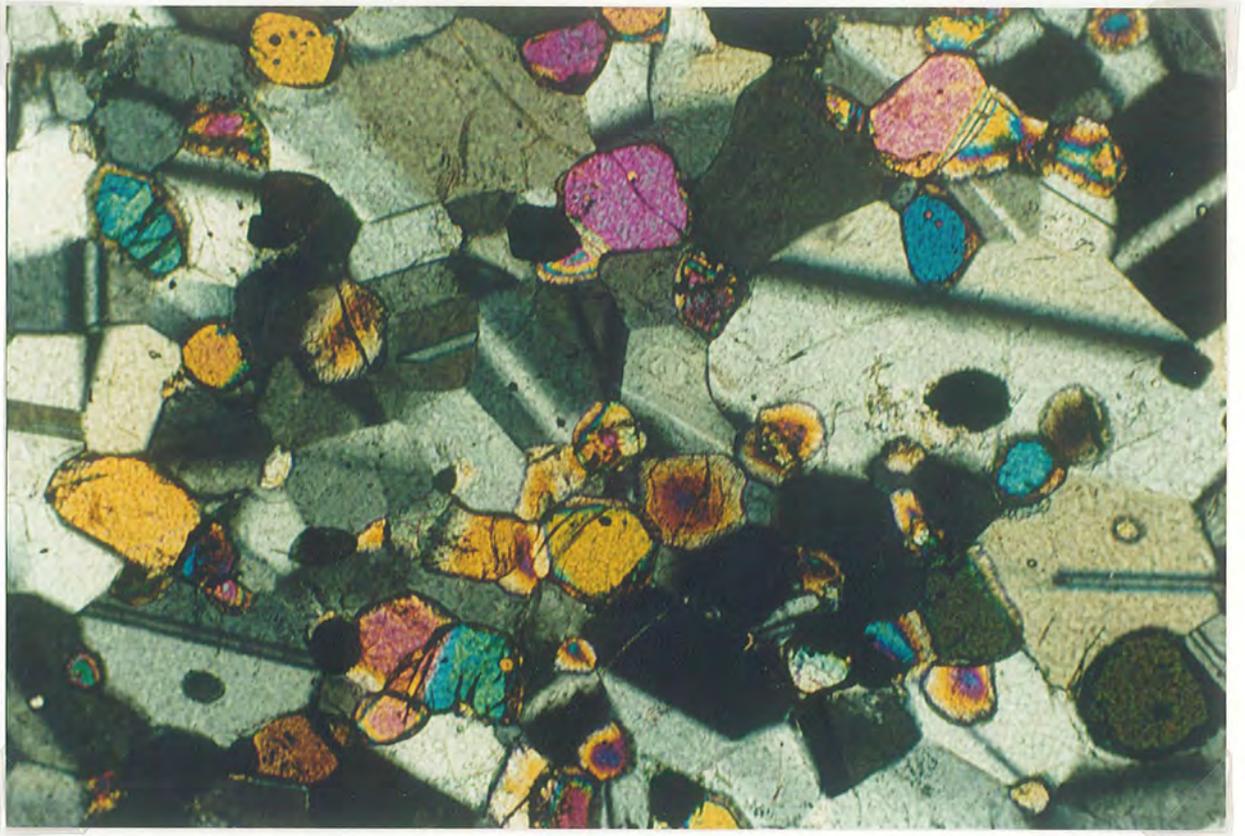
Fig. 3.29 Faults and Shear zones cutting the Inner Series of the Hypersthene Gabbro.



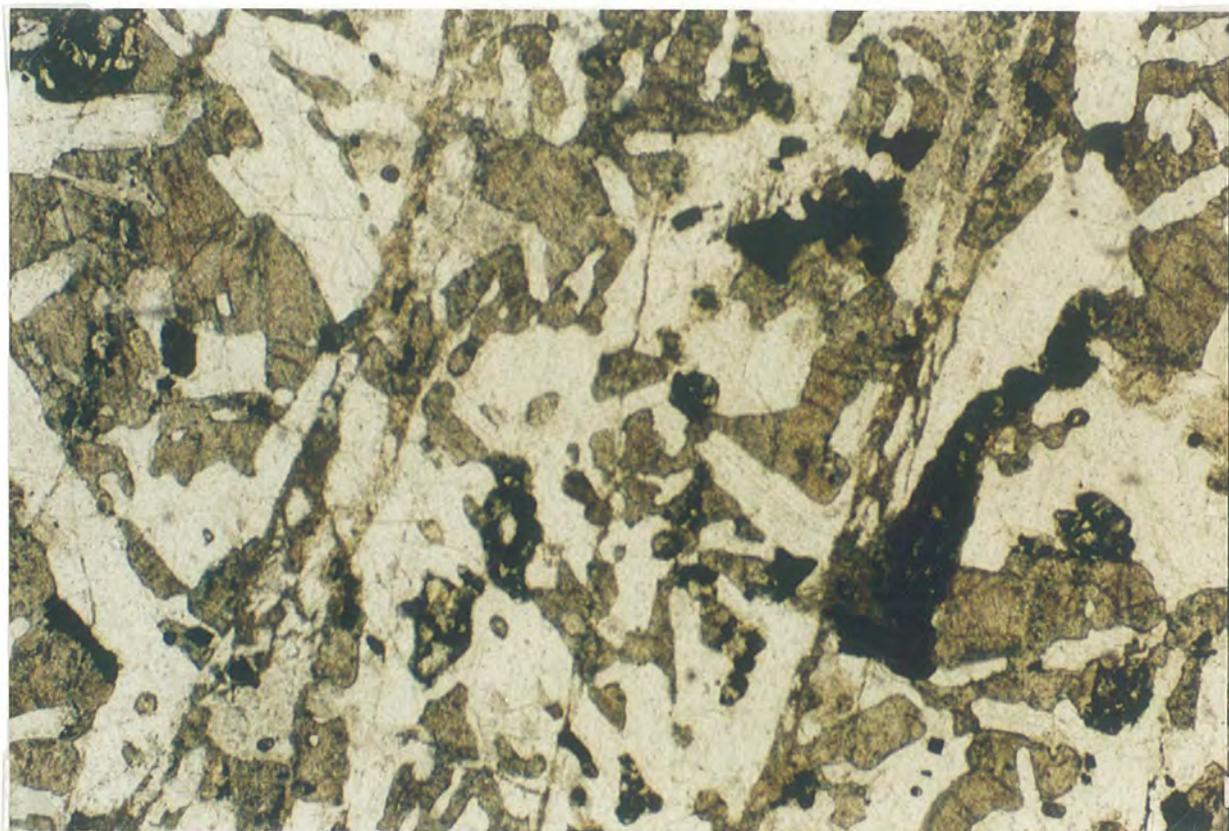
Glas Eilean fault active during most of history of complex.

- Exposed faults.
- g — Granulitic - textured high temperature shear zones.
- - - - - faults inferred from structural offsets.
- ⋯ Prominent linear gullies: may conceal faults eroded out as a result of hydrothermal weakening.
- ↘ Dip and offset on fault ( where known ).
- ↔ Strike - slip fault.
- ⊥ Vertical fault.

Plates 3.116A and 3.116B. Pristine pyroxene - hornfels metabasic rock from fault zone within Inner Series. Produced by annealing of fine - grained rock produced by ductile deformation at high temperature ( this is indicated by the strong crystallographic fabric apparent in Plate 3.116B, photographed with a 1 -  $\lambda$  plate inserted ). Note well - developed textural equilibrium apparent in Plate 3.116A. The rock ( sample 109/1 ) shows many similarities to sample 332 ( Plate 3.71 ). width of field of view 1.5mm in both.



**Plate 3.117.** Pyroxene - rich granular - textured microfault zones in fine - grained gabbro, sample 286/5. Plane - polarised light, width of field of view 3.5mm.



occur, particularly on the coasts. Mineral assemblages in these later faults and associated microfaults range from granular augite + orthopyroxene + opaques enclosing angular fragments of feldspar ( Plate 3.117 ) to chlorite - epidote - quartz rich hydrothermal vein assemblages. This suggests that activity on the faults continued long after the emplacement and most of the cooling of the Inner Series. Some of these faults, notably the fault running from the mouth of the Allt Grigadale ( Grid Ref. 422673 ), east of Ardnamurchan Point, to Reidh - Dhail ( 437644 approx. ), are cut by the next - oldest large intrusions in western Ardnamurchan. This is the most important field evidence for a period of relative inactivity between solidification of the Hypersthene Gabbro and emplacement of the later large intrusions.

The three groups of faults apparent in Fig. 3.29 appear to have been produced as follows:

- 1). Roughly concentric, steeply inward - dipping faults mostly, if not all, related to central uplift of the Inner Series. These include the reverse faults which bound the Inner Series on its southern side, and possibly on its northern side as well, and faults further within the IS. The last group of faults could however be associated with central subsidence since there are no sense - of - movement indicators or measurable offsets associated with them.
- 2). Faults striking  $140 - 150^\circ$ , which are subvertical to inward - dipping. These are parallel to the regional structural trend ( see Chapter 2 ) but are best developed on the western side of the Hypersthene Gabbro where these are tangential to the central uplift ( or subsidence ) and could represent pre - existing structural weaknesses exploited by this, or result from the interaction of local and regional stress fields ( see Chapter 7 ).
- 3). Compartmental or transfer faults, mainly radial faults, which transfer displacement between the two groups above, or between members of the same group of faults. They are commonest at the intersection of the two main groups of faults at An Acairseid.

As noted in section 3.4.1, the very highest - temperature, super - solidus deformation in the laminated gabbros and the gabbroonorites is pervasive and occurs throughout the Inner Series, except in the finer - grained, more doleritic - textured gabbroonorites. There is no obvious overall variation in the intensity of this deformation such as is required by the central subsidence model of Skelhorn & Elwell

( 1971 ). These workers measured the dips of a wide variety of structures which they considered to be primary igneous layering or lamination, and argued that there was a gradual increase in the inward dips shown by these structures towards the centre of the pluton, to angles of as much as  $60^\circ$  on Beinn na Seilg and at Sanna. They interpreted this as an inward - verging annular monocline which folded originally horizontal cumulate structures and was produced by tectonic subsidence, at high temperature, of the centre of the pluton. This implies an inward increase in shear strain, measured vertically, from about 0.17 at the periphery of the fold, where dips are about  $10^\circ$ , to 1.7 at the inner edge of the part of the pluton which has been preserved. No evidence, in the form of phenocryst or poikilocryst rotations, or the development of vertical fabrics, was found for pervasive shearing of this type. The discrete granular concentric shear zones described above, although perhaps associated in some cases with central subsidence, could not produce the rotation of the intervening rocks which is required by the Skelhorn & Elwell model.

Furthermore, the underlying assumption that the structures whose orientations were measured were originally horizontal is placed in doubt by the recognition that many of the planar fabrics used in Skelhorn and Elwell's work must have been internal intrusive contacts or occur in isolated xenoliths which may have been rotated during incorporation into the gabbro intrusions in which they now occur. Only measurements of lamination or layering in large, laterally extensive outcrops of laminated gabbro are at all plausible as valid estimates of post - formational tectonic rotation, and these are too few and far between to account for all of Skelhorn & Elwell's data points ( see maps in Skelhorn & Elwell 1971 ). A number of dip measurements satisfying this criterion were made during the course of the present work. The results suggest that the laminated gabbros do indeed show an overall inward increase in inward dip, but the number of possible measurements is small and the dips show strong local variation, particularly at the inner margin of the present exposures of the Inner Series. For example, the 4 dip measurements made around the summit of Beinn na Seilg range from  $25^\circ$  to  $40^\circ$ : when the data from 5 smaller, probably or definitely xenolithic sheets are included the range expands to  $20 - 90^\circ$ . The evidence for the Skelhorn & Elwell ( 1971 ) model is therefore much more ambiguous than was originally thought, and given the abundance of post - laminated - gabbro intrusions ( principally the gabbro intrusions ) could equally well be accounted for by displacements associated with these intrusions ( see Chapter 7 ): variations in the orientation and effects of different intrusions could account for the scatter in the data.

### 3.4.8. A summary of the geology of the Inner Series and its age relationships to adjacent rocks .

Evidence presented in sections 3.3 and 3.4 has shown that the Inner Series of the Hypersthene Gabbro is not a cumulate sequence in the classical sense of a sequence of rocks formed by progressive accumulation of a crystal - supported crystal mush from a single large body of magma, and modified by subsequent interstitial crystallisation and/or cumulate compaction processes. Instead, the numerous planar, shallowly dipping intrusive contacts within the gabbro-norites which make up the bulk of the Inner Series, the various cross - cutting relationships between these rocks and syn - magmatic faults in the An Acairseid area, and the presence of screens of older rocks between sheet intrusions, indicate that the Inner Series was built up over an extended period of time as a series of confluent, inward - dipping sheet intrusions emplaced into older rocks and earlier intrusions of the series whilst the latter still contained interstitial partial melt. The internal structure of the Inner Series as proposed here bears a close resemblance to the models of Wells ( 1978 ) and Walker ( 1975 ), but the external contact relationships proposed here are quite different. This is because of the recognition of the Marginal Border Group as a separate large, steep - sided intrusion and the discovery of steep faults separating the MBG and the IS along much of the boundary between them, in addition to nearly flat - lying intrusive contacts in some places. As shown in section 3.3, activity on the faults overlapped with emplacement of sheet intrusions within the Inner Series, suggesting that the faults were emplacement - related. They show evidence of central uplift in most cases and this would be an appropriate mechanism for creating space for the gabbro-norite sheets.

Emplacement of the gabbro-norite sheets into a thick sequence of partially molten rocks is indicated by the development of microstructures and textures similar to those produced by postcumulate processes in true cumulate sequences. These include deformation of early - formed crystals ( mainly plagioclase and olivine ) prior to crystallisation of interstitial phases ( mainly pyroxenes and opaques ), by crystal mush compaction which also produced secondary lamination and bending of the lamination above and below rigid poikilocrysts, and an absence of late - crystallising patches of residual melt, indicated by concentrations of phases such as quartz, alkali feldspars, apatite and zircon. The absence of these phases indicates the operation of an efficient mechanism of residual melt extraction. The development of metasomatic or replacive pyroxenites, olivine pyroxenites and pyroxene - rich gabbros, which cross - cut both the gabbro-norites and screens of older rocks, also requires large

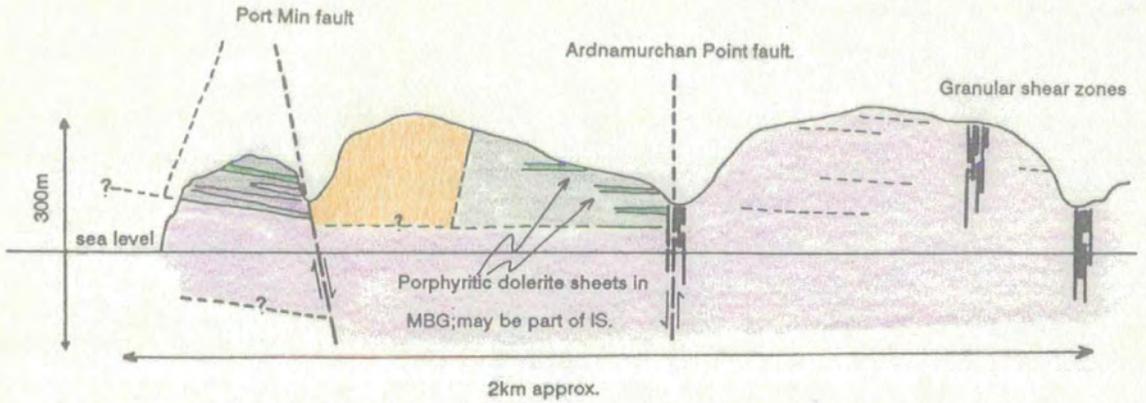
- scale interstitial melt mobility ( see Chapter 6 ) within the Inner Series. During the period of emplacement of the gabbronorites, the Inner Series appears to have formed a large body of hot, partially molten rock with, at any one time, at most only a few sheets of largely molten magma enclosed within it. It is of interest to note that many geophysical studies of active magma chambers beneath modern volcanoes indicate that they have a similar structure, with large regions of low seismic velocities and strong shear - wave attenuation, corresponding to rocks containing interstitial melt, and few if any regions of complete S - wave absorption, corresponding to bodies of magma with or without suspended crystals, but lacking a stress - transmitting interlocking network of crystals: see reviews by Iyer (1984 ) and Ryan ( 1987 ).

The geometry of the preserved part of the Inner Series close to the present level of exposure is summarised in Fig. 3.30. This consists of two sections because the Inner Series, unlike the Marginal Border Group, shows marked departures from radial symmetry, particularly in the form of its outer contact. The northern and southern contacts, except at An Acairseid, are steep and probably formed by a single large fault concentric to the pluton, along which central uplift probably took place. In contrast, the western margin consists of a series of shallowly - inclined intrusive contacts between gabbronorite sheets and the MBG, which have been offset by syn - emplacement movement on NNW - SSE trending linear fault zones. The latter are parallel to a tangent to the outer contact of the pluton in this area and also to regional structures. The complex zone at An Acairseid seems to be a transition or transfer zone between the two types of contact. The eastern margin of the Inner Series has not been preserved. The presence of both regional - trend structures on the one hand, and radial and concentric structures on the other, suggests either exploitation of pre - existing zones of weakness or the influence of both regional and magmatic stress fields on its emplacement: this is discussed further in chapter 7.

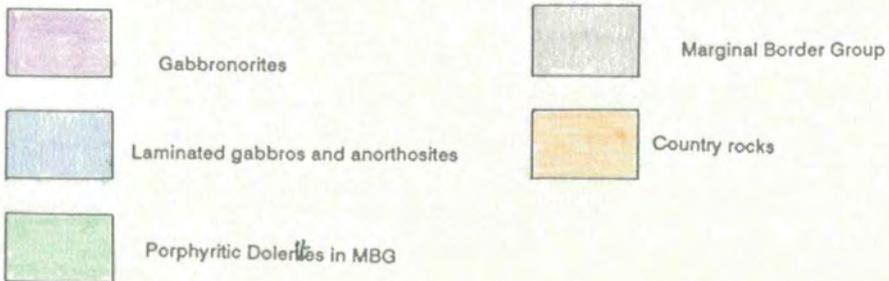
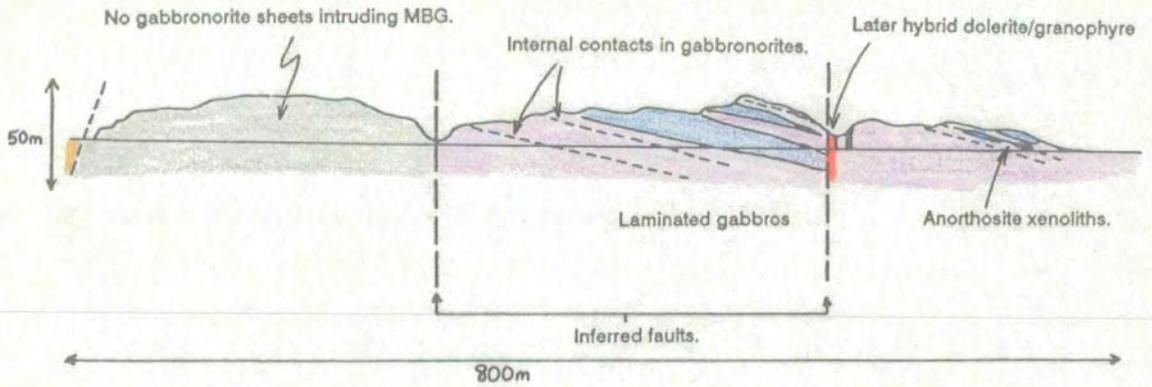
Apart from a few plagioclase porphyrocryst - rich layers in the gabbronorites which might have been produced by settling of the porphyrocrysts within sheet intrusions, the only rocks with a really close resemblance to cumulates within the Inner Series are to be found in the screens and xenoliths of older rocks which occur between and within the gabbronorite sheets. These are very varied rocks, and while the various laminated gabbros found around the Inner Series might all be part of a single earlier intrusion, the others have widely differing mineral assemblages and it seems unlikely that they could all form part of the same intrusion ( see Chapter 6 ). All must have been brought up to

Fig. 3.30 Sketch sections through the preserved parts of the Inner Series ( not to scale ).

A. ENE - WSW, through Gharblach Mhor and Druim na Cloise to Druim Reidh - Dhalach.



B. North - South, from Sanna Point to Sanna Bay.



their present positions during the emplacement of the gabbronorites ( see Chapter 7 for estimates of the distances involved ) and would have originally crystallised below the present level of exposure.

Subsequent to emplacement of the gabbronorites and the compaction in any one area, a group of porphyroblastic beerbachitic dolerite sheets were emplaced and produced localised partial melting of the enclosing gabbronorites and laminated gabbros. These beerbachites were emplaced during very high - grade hydrothermal metamorphism of the Inner Series, which post - dates the end of textural re - equilibration of all but the finest - scale textures in the rocks in which it occurs. The main effects of this metamorphism were oxidation of olivine in particular and the formation and plugging of tensile fracture networks, largely filled with anhydrous mineral assemblages. This metamorphism is best developed in the interior of the Inner Series: in the vicinity of intrusive contacts against Marginal Border Group rocks the IS rocks show only very slight, if any, developments of very high grade assemblages, and relatively intense development of subsequent more moderate - grade hydrous assemblages. Similar assemblages also partly overprint the earlier anhydrous igneous and metamorphic assemblages in the interior of the Inner Series. This later, lower grade metamorphism partly overlaps with the emplacement of a sparse group of orthopyroxene - free, intensely feldspar - phytic dolerite and quartz - dolerite intrusions. Even later chilled dolerite and mixed - magma sheet intrusions, which appear to be part of the same swarm of minor intrusions to the south of the Inner Series, are probably offshoots from later large plutons unrelated to the Hypersthene Gabbro.

Over small areas, from thin sections to groups of outcrops, the sequence of events outlined above is consistently adhered to, even if not all are present in the history of a particular section or outcrop. However, on the larger scale there is considerable overlap between them, particularly magmatic and structural phenomena such as the overlap of emplacement, crystallisation and compaction of gabbronorite sheets with faulting around An Acairseid. This is a natural consequence of the extended period over which the gabbronorite sheets were emplaced. As is the case with the MBG and its metamorphic aureole, perhaps the most reliable indicator of the passage of absolute time is likely to be changes in the composition of the magma supplied to the Inner Series intrusions. The change from orthopyroxene - rich olivine gabbronorites and orthopyroxene - bearing beerbachitic dolerites to the post - metamorphic orthopyroxene - absent dolerites suggests that such changes do occur. The consistent age relationships of these compositional groups to the metamorphism suggests also that the latter stages of this, at least, record slow continuous cooling of the Inner Series as a whole.

The age relationship of the Inner Series to the Marginal Border Group was established unequivocally in section 3.3: it is younger than the MBG. The question of how much younger, even in relative terms, is less easily answered. However, the lack of intense chilling at the outer intrusive margins of the Inner Series suggests that its emplacement began while the MBG and the adjacent country rocks were still at moderate metamorphic temperatures, although this may have been the result of reheating of the earlier rocks by precursor intrusions to the Inner Series which have not been preserved. The migration of high temperature hydrothermal fluids through the Inner Series must, for reasons to be discussed in Chapter 6, have been associated with lower - temperature hydrothermal activity in the surrounding rocks. These constraints suggest that the Inner Series was emplaced in the earlier part of M2 and that M2 continued until the Inner Series cooled. This in turn suggests that the late - M2 porphyritic dolerites ( Group 3 of section 3.2.5.4 ) outside the pluton and the late porphyritic dolerites within the pluton were emplaced at much the same time.

### 3.5. A summary of the geological history of the Hypersthene Gabbro.

Although many of the various events and episodes of activity associated with the evolution of the Hypersthene Gabbro as an active magma reservoir overlap with periods of different types of activity ( hence the need for correlation diagrams such as Fig. 3.5 ), there is an overall sequence of periods characterised by particular types of deformation, intrusion of different rock types or, most especially, different conditions of metamorphism. More or less the same sequence of such periods of activity is followed all around the pluton, although they were not necessarily always synchronous.

The earliest activity interpreted as being related to the Hypersthene Gabbro is the development of the dome structure around its present position during the latter part of the early hydrothermal metamorphism, M0. Although this doming, which is a form of central uplift of the country rocks, ended early on in the history of the pluton, central uplift continued intermittently throughout much of the history of the pluton, by movement on concentric inward- dipping reverse faults, on concentric thrust faults, and by emplacement of cone sheets. Like the doming, however, movements on large, steep, approximately concentric faults in the Glendrian Bay area ( and possibly on the Kilchoan Bay fault as well ) which also produced central uplift, seem to have ceased early in the history of the pluton, and certainly prior to the end of high grade metamorphism.

Emplacement of a large magma chamber corresponding to the Marginal Border Group resulted in the next period in the history of the Hypersthene Gabbro. This is characterised by a period of thermal contact metamorphism, M1, which produced a contact aureole 300 - 700m wide. Extremely high temperatures were attained close to the contact and partial melting of the country rocks occurred. Mobilisation of the least refractory lithologies produced rheomorphic breccia intrusions. However, large - scale mixing of country rock melts with the intermediate to basaltic magmas within the magma chamber appears to have only taken place in and around discrete wedges of country rocks projecting into the magma chamber. These were produced by downfaulting on a system of concentric, inward - dipping extensional faults linked by radial faults which were also active at the time. This faulting produced central subsidence around and above the magma chamber and must have alternated with central uplift by movement on the reverse faults noted above. The high - degree partial melts produced in the downfaulted wedges, together with granitic and granodioritic magmas also present

at the wall of the magma chamber, were involved in multistage magma mixing with more basic magmas. This produced, along parts of the outer contact of the Marginal Border Group, a highly complex and varied contact zone suite. Incorporation of refractory material from the country rocks at this time produced xenolith swarms. Some early cone sheets may also have been emplaced at this time.

M1 ended with the abrupt quenching of the contact aureole, and the contact zone of the magma chamber in places, followed by slower but still rapid cooling of the remainder of the MBG as presently preserved. This quenching event ( or events, in different parts of the aureole ) was associated with the formation of a tensile fracture network and was in places rapid enough to quench felsic melts to glasses. Rapid solidification of the contact zone as a result of this quenching event appears to have resulted in the preservation of many of the liquid - liquid interaction structures in the contact zone suite. Subsequent to the quenching event and the onset of hydrothermal ( M2 ) metamorphism produced by fluids circulating along the fractures, **at the present level of exposure**, numerous bodies of granodioritic magma were emplaced in the country rocks close to the contact. These were quenched to felsites for the most part, which resulted in the preservation of relict porphyrocrysts and microxenoliths in the felsites, suggesting that they may have been produced by continued melting at the wall of the magma chamber at depth. The M2 metamorphism caused partial overprinting of the M1 aureole and also defines a zoned aureole centred on the Hypersthene Gabbro. The earliest parts of M2 in the inner parts of this aureole are prograde in places but in general the preserved M2 assemblages record progressive retrograde metamorphism. The majority of the cone sheets around the Hypersthene Gabbro appear to have been emplaced during the earlier part of M2 and contain early M2 assemblages. Central subsidence on some of the concentric inward - dipping faults continued and alternated with movements on concentric thrust and reverse faults which produced central uplift. Later in M2, when the overall grade of metamorphism had declined considerably, a distinctive suite of strongly feldspar - phyric dolerite cone sheets were intruded, as were various small gabbros in the Glebe Hill area.

Emplacement of the Inner Series of the pluton probably began during the earlier part of M2, although after emplacement of the petrographically distinct main group ( Group 2 ) of cone sheets in the country rocks. The Inner Series is dominated by a thick sequence of confluent gabbronorite sheet intrusions which were emplaced in rapid succession, producing a large body of partially molten rock

which underwent compaction and residual melt expulsion. Emplacement of this body at its present level of exposure was largely accomplished by central uplift on syn - magmatic high - temperature shear zones, although gabbro-norite sheets were intruded into the adjacent country rocks during the same period. After the gabbro-norites were emplaced at any one outcrop a weak to locally strong lamination developed in them by a process of compaction and recrystallisation. The formation of the lamination may, however, have overlapped with the emplacement of gabbro-norite sheets in adjacent parts of the Inner Series as it must have taken place whilst the Inner Series as a whole was still partially molten. In addition, replacive or metasomatic pyroxenites formed in a few places. Following this period of recrystallisation as developed at any one locality, tensile fractures began to form and high grade, pyroxene - hornfels facies, hydrothermal circulation and metamorphism took place. During this period of metamorphism, beerbachitic dolerite sheets and veins were intruded and produced very localised melting of the host rocks. Grades of metamorphism in the Inner Series declined thereafter and partial overprinting by hydrous mineral assemblages took place. There is no good evidence for folding of the Inner Series by central subsidence at this or any other time.

After cooling of the Inner Series to, at most, hornblende - hornfels facies conditions, a variety of compositionally distinct, orthopyroxene - poor minor intrusions were emplaced. Some of these appear to be related to later large intrusions whilst others may be correlated with later cone sheets outside the Inner Series and gabbros in the Glebe Hill area.

As noted in section 3.2.7 with reference to the Marginal Border Group and the contact aureole, the reconstruction of the outline history above depends on correlating characteristic events between different areas which are separated by areas of poor or no exposure. The principal means of correlation has used the phases of metamorphism as age indicators because these can be relied upon to affect, to greater or lesser extents, all the rocks older than themselves within the areas in which they are developed. A second means of correlation used is to use compositionally distinct rock types as time indicators, and to an extent the two can be used to cross - check each other. However, neither can be considered *a priori* to not be diachronous around the pluton: demonstrations that they are or are not synchronous depend on an understanding of the processes involved in metamorphism or in the formation of the magmas which they represent, respectively. A number of possible indicators of diachroneity of metamorphism have been noted in this chapter, for example the different widths of the M1 aureole on the northern and southern margins of the pluton and the presence of apparently

anatectic melt intrusions ( the M2 felsites ) emplaced after the end of melting at the present level of exposure. The overlap of periods of emplacement of magmas of distinctive composition, in and around zoned magma bodies, has previously been demonstrated in other complexes ( for example, in the Rhum complex ( Greenwood 1987 ), where the magmas concerned may actually belong to entirely distinct magma series ). The question of whether such a situation arises in the Hypersthene Gabbro is considered in Chapters 4 and 6.

In one sense the phases of metamorphism must necessarily be diachronous, because they are thermal or hydrothermal in character, and the physical laws of heat transfer require that heating and cooling of large volumes of rock must take long periods of time during which temperatures will vary widely within those bodies of rock. The exact lengths of these periods, and hence the extent of diachroneity, is dependent on the particular heat transfer mechanisms and initial conditions involved ( see, for example, Eckert & Drake ( 1987 ); also Lister 1974, McKenzie & Bickle 1987; and the discussion of heat transfer around the Hypersthene Gabbro in Chapter 5 ). Conversely, the physics of heat transfer processes, given some knowledge of the situation in which they occurred, can be used to constrain the amount of diachroneity involved. A greater degree of uncertainty is introduced by the question of whether, say, the cooling recorded in one rock was produced by the same **geological** event as that recorded in others in a different area. One example, which will be considered in subsequent chapters, is the question of whether the quenching at the end of M1 was triggered by a single event all around the pluton or by successive events in different areas. Examination of problems of this type require an understanding of the mechanisms of the phenomena involved, and it is these which are examined in some of the subsequent chapters.

