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**GRANITOID EMPLACEMENT AND DEFORMATION: A CASE STUDY OF
THE THORR PLUTON, IRELAND, WITH CONTRASTING EXAMPLES FROM
SCOTLAND.**

by

Michèle A. Mc Erlean

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**A thesis submitted in partial fulfilment of the degree of Doctor of Philosophy at
the Department of Geological Sciences, University of Durham.**

1993



i

28 MAR 1994

To
MUM & DAD



The Bloody Foreland from Drumnacart

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Abstract

The importance of pre-existing country rock structure in controlling the siting and geometry of intrusive bodies has been overlooked by recent workers, with most studies tending to concentrate on the effect of tectonic processes. This thesis describes three Caledonian plutons emplaced at intermediate crustal levels, the Thorr Granite, Co. Donegal, the Ratagain Complex, Inverness-shire and the Loch Loyal Syenite Complex, Sutherland. None of these plutons appear to have been emplaced in association with large tectonic strains. In all three cases, data collected from both the country rocks and the plutons demonstrates that regional tectonics alone cannot account for the observed shape, fabric evolution and mode of emplacement. Instead these features are more directly controlled by the interaction of the pre-existing structural architecture in the country rocks with fault and shear zones. Thus, whilst tectonic forces act as the catalyst in initiating the creation of space into which magma can be emplaced, the pre-existing structural architecture will control the ultimate form of the pluton. The plutons described in this thesis can, therefore, be considered to represent an intermediate stage between the sheeted intrusions that are emplaced in association with active shear zones, for example the Main Donegal Granite, and high level, passively emplaced plutons, such as the Rosses Granite.

The evolution of deformation fabrics within the three plutons is described in detail. In all cases magmatic state deformation fabrics are predominant. However, most of these magmatic state fabrics are aligned parallel to the intrusion margins; this is particularly true in the case of the Thorr Pluton. Such a geometry could be accounted for by invoking the presence of large shear strains at the time of fabric formation. However, in the absence of evidence for large shear strains in the country rocks, it is proposed that such fabric geometries may be produced as a result of localised coaxial strain component in response to the body forces, or 'buoyancy head', exerted by the magma acting across the intrusion walls.

Finally, in studying the kinematics of intrusion of the Thorr Pluton, a new technique for determining shear sense in rocks deformed in the magmatic state has been applied. The data collected from the application of this technique was found to corroborate the shear sense data collected from the envelope rocks. In this instance, the technique was a valuable aid to kinematic analysis, and ultimately to deducing an emplacement model for the Thorr Pluton.

Acknowledgements

During the four years prior to the presentation of this thesis have marked one of the more challenging periods of my life. Many friends have been made along the way, people who, unlike many others, have tolerated my frankness, had the courage to draw their own conclusions, avoiding hearsay. To those people I owe a debt of gratitude.

My first thanks goes to Dave Hirst who took on the onerous task of supervising this work at a fairly late stage. Your time and consideration were greatly appreciated.

Dr. D.H.W. Hutton is acknowledged for introducing me to the subject of granite emplacement and deformation and the Thorr and Ratagain Plutons. It was unfortunate that he chose to put his personal opinions before his professional judgement; the rest, as they say, is history.

The Baikie family, Maureen, Derick, Tanya, Anthony and Michael, are thanked most sincerely for the warmest of Irish hospitality; they welcomed myself and my friends as part of the family.

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Finally, to Bob I owe more thanks than words can express. He has shown bravery, tolerance and support beyond the call of duty during the preparation of this thesis, so for him I'll finish by saying

"long live the conceited b*****s club."

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Part I

OVERVIEW

Chapter 1

GRANITOID EMPLACEMENT & DEFORMATION

1.1 Introduction

A fundamental way in which material is added to the continental crust is by granite emplacement (Wyllie *et al.* 1976). Granitic rocks are traditionally associated with zones of plate tectonic activity, which can be related to deformation at continental margins including settings such as subduction zones, continental collision zones and transcurrent and crustal extension zones. Granite *sensu lato* defines a wide variety of chemical and petrological compositions, and reflects the variation in the origin and source of melts as mentioned above. An important division is that between 'true granites', that is the crustal melts or migmatites of continental collision zone affinity and the voluminous tonalitic associations of the subduction related granitoids, both of which Read (1957) and Buddington (1959) incorporated in the 'granite series'. This classical approach related emplacement style to crustal level and resulted in the subdivision of plutons and batholiths into *epizonal*, *mesozonal* and *catazonal* categories in order of increasing depth. However, this classification was criticised by Pitcher & Berger (1972) who noted that different emplacement styles existed at the same crustal level in Donegal. It was after this important revelation that geologists began to realise that granite type and emplacement style could be generally related to plate tectonics and more specifically to the detailed structural setting. This approach was developed by Pitcher (1978, 1979) and has continued through the work of many geologists until the present day, when there is a strong realisation that the generation, ascent and emplacement of granitic magma are all fundamentally related to tectonic activity (Karlstrom 1989, Paterson 1989, D'Lemos *et al.* 1992, Hutton 1992).

There is great diversity in the shape and size of granite plutons and they can display a wide range of structural variation. The range of structures (including fabrics and foliations) encountered during a field study is likely to reflect the emplacement mechanism, except in rare situations when it may yield some information about the ascent of the magma from its source. Generally structures preserved are a response of the granitic magma to a combination of body forces, external tectonic forces and internal magma buoyancy. The overall response of the granite is controlled by the physical

properties of the magma which can determine the ability to record deformation. Particular magma chamber processes, such as convectional overturn of magma, would tend to destroy that record, but since the physical properties of the magma also controls the operation of such processes (M^cBirney & Murase 1984), then magmas with specific rheologies will tend to preserve deformation effects, whilst others will tend to lose them.

1.2 The Rheological Properties of Granite

Rheology is defined as "the science of flow of matter"; as such it is concerned with the properties of elasticity, viscosity and flow which materials exhibit when subjected to stress. The rheological response of a rock may vary according to changes in composition, but also with stress, strain and strain rate. Ideally, materials exhibit three main types of behaviour when subjected to an external stress field.

Elastic Behaviour:

Where stress (σ) is proportional to strain (ϵ) which is totally recoverable on removal of stress.

$$\sigma \propto \epsilon \quad [1.1]$$

This type of behaviour can be defined by Hooke's law, where:

$$\epsilon = \sigma/E \quad [1.2]$$

defined by Young's modulus, which is a measure of the material's resistance to elastic distortion,

$$E = \sigma / \epsilon \quad [1.3]$$

Ideal Plastic behaviour or Yield:

Rocks will tend to flow when they are stressed beyond a critical value called the Yield stress (Fig. 1.1) (σ_0). Deformation below this critical value is elastic and is, therefore, recoverable when stress is removed. Yield stress is not, therefore, a threshold value below which materials will not suffer deformation, but a point at which the nature of deformation transgresses from being non-permanent to permanent. However, a note of caution must be added at this point since all experimental data to date has been carried out at relatively fast strain rates. M^cBirney & Murase (1984) point out that at much slower strain rates, rocks may deform permanently below their yield strength by the process of creep.

This is something that we understand very little about, but which probably occurs naturally in most granitic rocks over long periods of time. As such, the yield strength of magma is more properly described as a transition value below which the dominant mechanism of deformation is not viscous flow, but another much slower process which is still capable of producing permanent strain.

Ideal viscous flow:

Viscosity is, in general, the coefficient for transfer of momentum (M^C Birney & Murase 1984) as defined by the equation

$$\tau = \tau_0 + \eta (du/dy)^n \quad [1.5]$$

in which τ is the shear stress applied in the direction x parallel to the flow plane, τ_0 is the minimum stress required to induce permanent deformation, du/dy is the velocity gradient normal to the shear plane, and n is a constant with a value of one or less. In the simplest situation, $n = 1$; this is known as Newtonian viscous flow. In this case, stress is proportional to strain rate ($\dot{\epsilon}^0$) as defined by:

$$\sigma \propto \dot{\epsilon}^0 \quad [1.6]$$

or

$$\tau = \eta \dot{\epsilon}^0 \quad [1.7]$$

and the fluid has no yield stress ($\tau_0 = 0$) (M^C Birney & Murase 1984) since the strain response is instantaneous and permanent. Most pure fluids exhibit Newtonian behaviour and magma has been modelled as such for convenience (Shaw 1965). However, in reality most magmas are not Newtonian, rather they exhibit more complex non-Newtonian behaviour typical of other complex fluids, such as blood, mud and paste, where the relationship between stress and strain rate is non-linear.

More complex rheologies:

These more complex rheologies are modelled by combining components of the three ideal types (Fig. 1.1). The most important in the present context is elastoplastic behaviour, in which the initial behaviour of the stressed material is elastic followed by permanent ductile deformation beyond the yield strength. It is quite common for this type of behaviour to be described as ideal plastic or yield behaviour, but, the author believes that they can be distinguished based on their deformation path following the point of yield. In the case of ideal plastic behaviour, the strain path subsequent to yield is Newtonian (i.e. ideal viscous flow), whereas, with elastoplastic behaviour the response is non-Newtonian, since strain hardening and strain softening can occur. Materials that exhibit elastoplastic

behaviour are known as Bingham bodies, and several authors, notably Pitcher (1987) have suggested that they are a good representation of how magmas, particularly granites, respond to external stresses. Under certain conditions, granites seem to display other rheological responses, such as dilatant fluid behaviour, where the granite can exhibit essentially brittle behaviour when intruded rapidly by syn-plutonic basic magma, but show stress relaxation through time, seen in the form of crenulate margins between acid and basic material (Pitcher 1987).

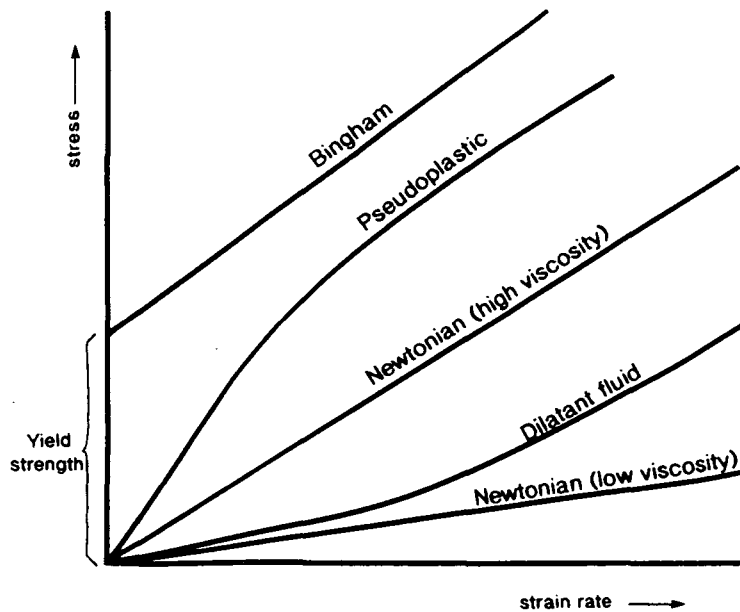


Figure 1.1 Graph of stress against strain rate to illustrate different types of rheological behaviour.

Whilst granitic magmas are likely to display a range of rheological properties during crystallisation and cooling, modelling experiments tend to rely on the simplistic assumption that their behaviour is Newtonian (van der Molen & Paterson 1979). Under such a regime, viscosity is governed by composition and temperature and is strongly dependent on volatile and halogen content in addition to strain and strain rate. If, however, the magma contains crystals in suspension, the crystal to viscous liquid ratio is an influential factor in determining viscosity; at low crystal content (< 50%) the behaviour of a magma will tend to approximate more closely to Newtonian viscous flow than at higher crystal content. However, the extent to which the behaviour remains Newtonian is also dependent on the size, shape and distribution of crystals within the magma (M^C Birney & Murase 1984), and also on magma composition.

Current models for the behaviour of magmas with increasing crystal content are based on the experimental work of Arzi (1978) and van der Molen & Paterson (1979) which suggests that there is a rapid, exponential increase in viscosity due to increasing interaction between grains. When the crystal content of the magma reaches a critical value, termed the Rheologically Critical Melt Percentage (RCMP) (Arzi 1978; Van der Molen & Paterson 1979) there is a marked change in the rate of increasing viscosity, which is shown by the pronounced break in the slope of the graph (Fig. 1.2). In real terms, this is the point at which the behavioural response is basically that of a solid, although each crystal may still be surrounded by a thin melt film which is capable of transmitting stress in a similar way to a solid material. Van der Molen & Paterson (1979) estimated that the RCMP has a value of approximately 30-35 %, which correlates well with the results from soil mechanics, melting experiments in garnet lherzolite, viscosity measurements of dense suspensions and phenocryst concentrations found in dykes.

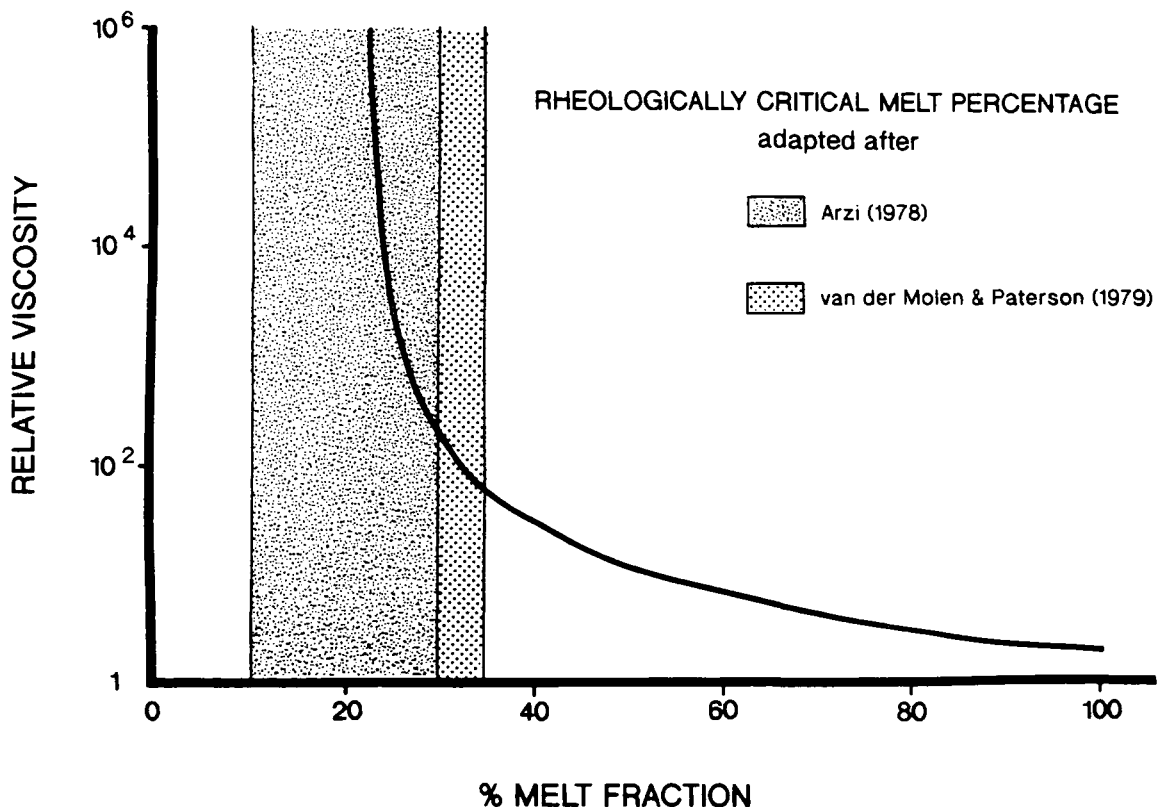


Figure 1.2 Graph of relative viscosity against strain rate. The exponential form of the curve illustrates the abrupt change in relative viscosity between 20%-35% remaining melt fraction; this is the rheologically critical melt percentage (RCMP).

Below the RCMP, the behaviour of magmas is increasingly dependant on the ability of the solid framework to deform plastically. According to Tullis & Yund (1980), a likely deformation mechanism under such conditions is dislocation creep or 'power law creep', which involves the intracrystalline movement of dislocations.

Yield strengths have been measured for Hawaiian lavas (Shaw *et al.* 1968), but no satisfactory measurements have been performed in granitic melts up to the present. M^C Birney & Murase (1984) pointed out that there are uncertainties surrounding the absolute importance of silica content to the behaviour of melts at the lower temperature ranges at which granitic melts form. Thus there is a fundamental question as to whether the yield strength is directly related to the behavioural response of the crystallising matrix, or whether it is related to a property which is, as yet, unidentified, perhaps silica bond strength? A further problem is that the limited experimental modelling that has occurred has been restricted to silicic lavas such as rhyolites and andesites (M^C Birney & Murase 1984) and has not been more widely applied to tonalites, granodiorites and diorites. In studying the rheology of these compositions we need to consider field observations to give us qualitative information about their behaviour during crystallisation and cooling.

1.3 The development of tectonic structures in granitic rocks

Structures in granites, as discussed in this thesis, include fabrics and foliations, contacts, shear sense indicators, mylonite zones, brittle faults, fault breccia and gouge. Current philosophy suggests that the crystal content of a magma at the time of deformation can be qualitatively estimated with reference to the types of structures observed in hand specimen and thin section. The development of such structures in granitic rocks must be rheologically feasible. However, simplified idealised models cannot usually be applied in a definitive manner, given that magmas do not usually appear to conform to any single type of rheological behaviour (as suggested by the experimental data of M^C Birney & Murase 1984).

Deformation of granitic magmas has been the subject of much attention during the past decade. Key works on the subject include Blumenfeld & Bouchez (1988), Hutton (1988a), and Paterson *et al.* (1989). The various authors have defined their own terminology for the identification

and discussion of fabrics in granitic rocks, all of which are broadly synonymous, and all of which define the structural age of fabrics relative to the crystallisation state of the magma. However, the diversity of definitions can be confusing to the reader, so the following is an attempt to outline the terminology as it will be used in this thesis and cross-reference it with synonymous terms:

1.3.1 *Magmatic state fabrics*

This describes a fabric which has developed while the granite is above the RCMP (Blumenfeld & Bouchez 1988). The development of such a fabric involves essentially rigid body rotations of early formed phenocryst phases such as biotites and feldspars. Consequentially, magmatic state deformation fabrics are characterised by a shape preferred orientation of the early phenocrysts, set in a groundmass of undeformed later phases, such as quartz (Fig. 1.3a). Hutton (1988a) called such fabrics 'Pre-full crystallisation' (PFC) fabrics. Paterson *et al.* (1989) called these 'magmatic flow' fabrics, but, in the opinion of this author, the term flow can cause problems due to a perception that the process of flow is initiated by internal body forces, whereas many, although not all, of these fabrics are produced with application of an external tectonic force on the magma body. Paterson *et al.* (1989) also had a subdivision called 'submagmatic flow', which according to this author's interpretation, characterises fabric development at, or slightly above the RCMP, where suspension-like behaviour starts to break down because of the impingement or collision of phenocrysts with one another. This work will describe these features as 'high temperature solid state' fabrics.

1.3.2 *Solid state fabrics*

Below the RCMP, even though there may be up to 30% melt remaining, the behaviour of the granite is essentially that of a solid, since an interconnected lattice structure is formed by the crystal framework and the surrounding melt occurs as thin films which are capable of transmitting strain. As a result, the crystal lattice of both phenocryst and groundmass minerals may be distorted (Fig. 1.3b). Quartz is particularly susceptible to distortion and recrystallisation during solid state deformation, so it has been widely studied in order to document the development of intracrystalline structures, especially under conditions of decreasing temperature. Studies have shown that quartz aggregates are relatively easily flattened or squashed compared to feldspar and mafic minerals, so they tend to

become elongated and will eventually undergo recovery and/or recrystallisation forming subgrains and new grains respectively, resulting in the development of lattice preferred orientations (White 1977). The end product of a continuous down temperature solid state deformation is mylonite if the deformation remains ductile. Paterson *et al.* (1989) subdivided solid state fabrics into high temperature and low temperature types. Hutton (1988a) called these fabrics 'crystal plastic strain' (CPS) fabrics, however, this can be misleading since the deformation mechanisms involved are not confined to crystal plasticity, but may include mechanisms such as grain boundary sliding, Coble creep or pressure solution (c.f. Knipe 1989).

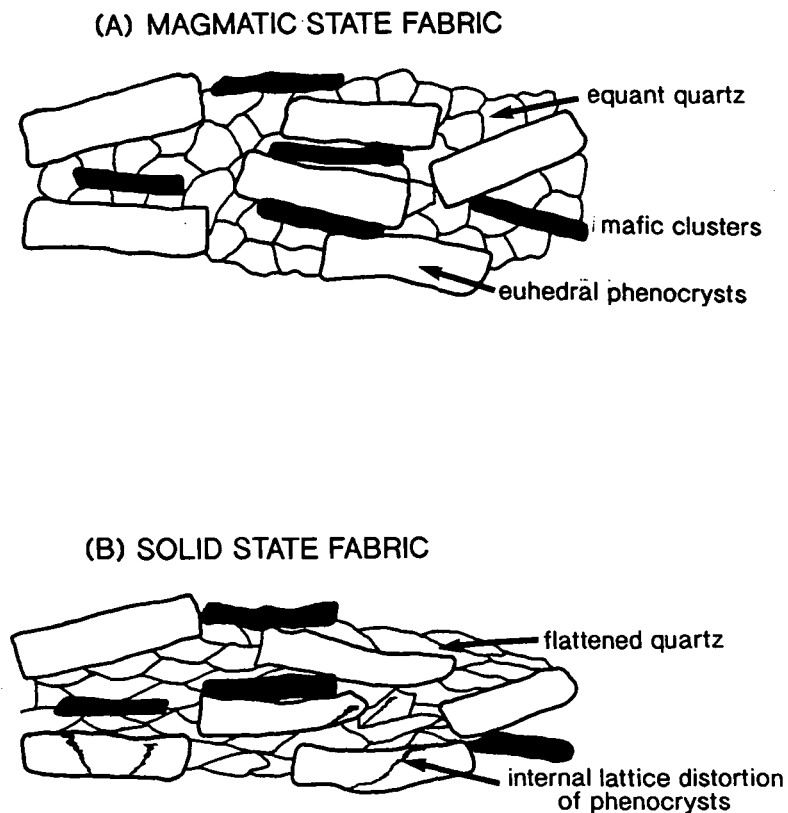


Figure 1.3 Characteristics of magmatic state and solid state deformation fabrics (adapted after Hutton 1988).

Obviously terms such as 'magmatic state' and 'solid state' fabric should be regarded as representing end-member situations, with the reality being an array of intermediate and often overprinting fabrics, with a variation in characteristics due to differences in the behaviour of magmas with slightly different compositions, volatile content and so forth.

1.3.3 *The brittle-ductile transition*

In quartzo-feldspathic rocks, such as granitoids, the transition from truly brittle to wholly ductile behaviour encompasses a wide range of physical conditions (i.e. temperature, strain rate, fluid pressure, confining pressure, etc.). Indeed one must exercise great care when ascribing the terms 'brittle' and 'ductile' since the bulk behaviour can vary according to the scale of observation, for example cataclastic zones can exhibit ductile flow processes in bulk terms (see Rutter & Brodie 1991). In general, however, the passage between greenschist and epidote-amphibolite facies is regarded as the major transition from a microstructural viewpoint. The most striking change is in the behaviour of feldspars, which exhibits mainly brittle deformation features, including brittle grain size reduction under greenschist facies conditions, whereas the amphibolite facies is dominated by ductile processes such as dynamic recovery and recrystallisation, which produces features such as ribboning of quartz (c.f. Mitra 1978, 1984; Simpson 1985; Gapais 1987, 1989). This change in microstructural character is also associated with a change in the distribution of fabric development to an increasingly heterogeneous distribution at lower temperatures.

1.3.4 *Strain Analysis*

It is usual to classify magmatic state and solid state fabrics as planar, linear or a combination of both, and to relate their fabric components to the strain ellipsoid using the nomenclature of Flinn (1965). This records the relative amount of longitudinal strain in three orthogonal directions of a hypothetical initial sphere, including the directions of maximum finite extension (X) and shortening (Z). Several ellipsoid shapes are possible depending on the longitudinal strain suffered by the third and intermediate strain axis (Y). S-type planar fabrics are associated with flattening or oblate strains, where the Y axis is extended by an equal amount as X. L-type fabrics are formed by constrictional or prolate strains, where the Y axis is shortened by the same amount as Z. LS-type fabrics, which occur

more commonly than the others, are intermediate between the two end members and include plane strains where the Y axis remains unchanged in length (Fig. 1.4). The nature of the fabric in many plutons can initially be determined by qualitatively estimating the degree of preferred orientation of the constitutive minerals. However, it is useful to quantify this in terms of the K-value of the strain ellipsoid, where

$$K = (X/Y - 1) / (Y/Z - 1) \quad [1.8]$$

in order to map out strain gradients within a pluton. This can be achieved by the measurement of the shape ratios of enclaves contained within the pluton. Cognate xenoliths are regarded as being most suitable, since they are assumed to have had minimal ductility contrast with the host granitoid at the time of deformation. Moreover, they are believed by some to approximate best to an originally spherical shape (Hutton 1988b), but this is not always the case, so it is important to realise that strain analysis using this method may over- or under-estimate the true amount of strain, depending on the actual initial shape of the inclusion. The technique involves measurement of the relative length to width ratios of the enclaves in the foliation plane (XY) and also in the plane orthogonal to the foliation and lineation (YZ), these values can be plotted directly onto the Flinn diagram. The K-value can then be determined by finding the slope of the best-fit line through all the data. It is also possible to calculate the ratio in the third dimension (XZ), according to the equation

$$R_{XZ} = R_{XY} \cdot R_{YZ} \quad [1.9]$$

this is an absolute measurement of the magnitude of the distortion at a specific locality. It can also be qualitatively determined from the Flinn plot, since the further away from the origin a point falls, the greater its value of R_{XZ} and the more distortion it has suffered.

In addition, the shape of the strain ellipsoid may be estimated using the 'Fry Method' (Hanna & Fry 1979, Fry 1979) which makes use of the deformed distribution of phenocrysts and can be carried out in the field or laboratory. This method also has limitations, since it requires a more-or-less homogeneous distribution of phenocrysts before the initiation of deformation, otherwise, the distribution after deformation retains a component of the original non-random distribution (see discussion in Ramsay & Huber 1983).

The Flinn Diagram

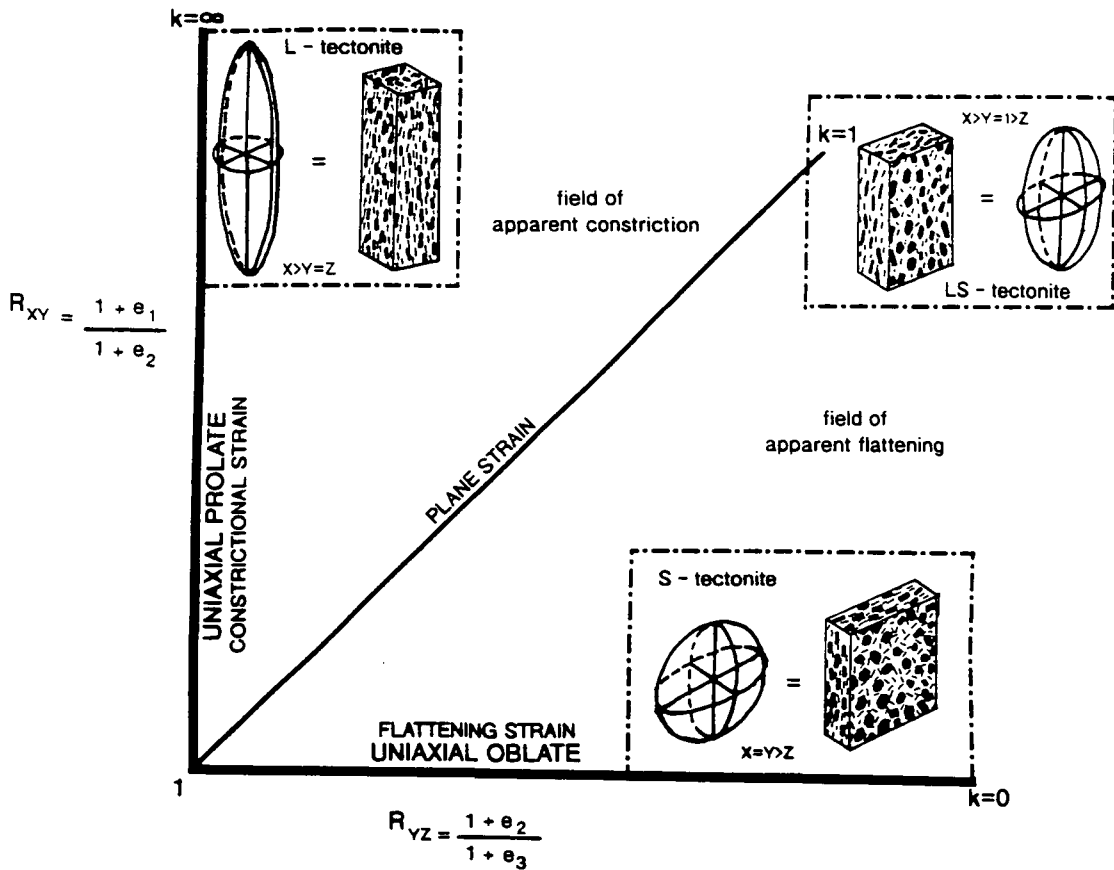


Figure 1.4 The Flinn Plot for determination of the distortional component of strain. Diagram also shows the principal strain field and end member tectonites associated with them.

It is important to realise that both the Fry and Flinn plot methods record only the distortional component of deformation. If the strain is non-coaxial (see below) the rotational component is very difficult to quantify, or even estimate, unless simple shear is assumed (See Ramsay & Huber 1983).

1.3.5 Shear sense criteria

The emplacement of granites in zones of active tectonism results in the magmas being subjected to the type of deformation, coaxial or non-coaxial, that is associated with that zone. As a general rule, non-coaxial deformation tends to occur more commonly than coaxial deformation, although the process of 'ballooning' in some plutons produces mainly flattening strains and hence S-

type fabrics (see section 1.4). As with non-coaxial deformations in other rock types, asymmetries are produced in the granitic structures, both meso- and microscopically and these can be used as shear sense indicators. It is obviously important to make observations about shear sense if one is to understand the kinematics of the deformation event. In order to determine the true sense of shear, the indicators must be viewed perpendicular to the main foliation and parallel with the transport direction or stretching lineation (Fig. 1.5). Shear sense can be determined from both magmatic and solid state structures, although they are better known in the solid state.

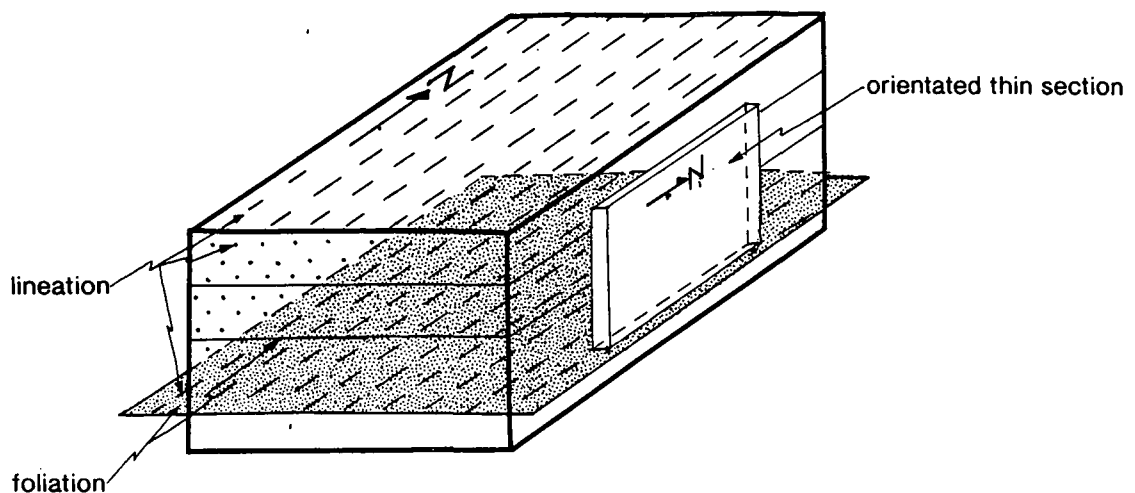


Figure 1.5 Cartoon sketch to illustrate the position of orientated thin sections relative to tectonic foliation and lineation.

1.3.5.a *Sense of shear in the magmatic state*

This is a topical subject presently since new methods have been developed in the past few years. Blumenfeld & Bouchez (1988) summarised the classical magmatic state shear sense indicators (Fig. 1.6) as:

- (1) obliquity between intrusion walls and a planar foliation formed by early phenocrysts (Fig 1.6.a);
- (2) sense of tiling of megacrysts (requires intermediate to high megacryst content) (Fig. 1.6.b)

(3) obliquity between subfabrics (either shape dimensional or crystallographically) (Fig. 1.6.c; see also Chapter 3).

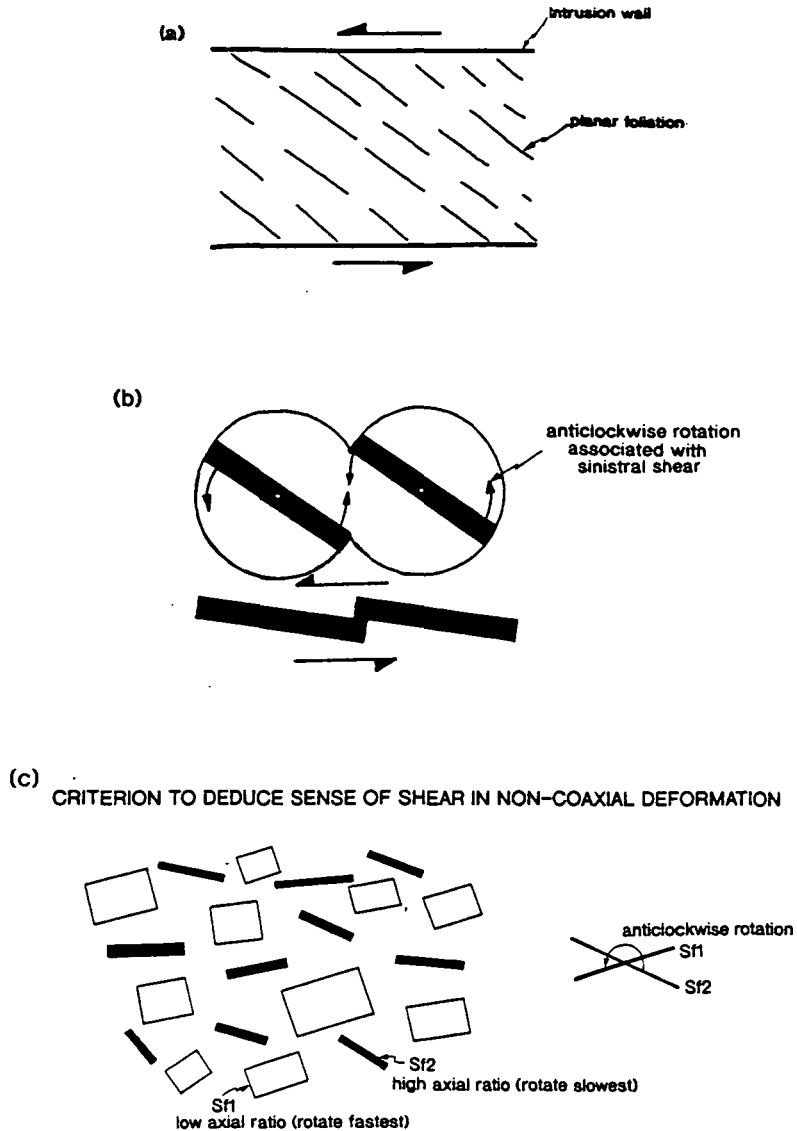


Figure 1.6 Criteria to deduce sense of shear from rocks deformed in the magmatic state in a sinistral shear regime. (a) obliquity between planar foliations and intrusion walls; (b) rotation and collision of phenocrysts to form tiling fabrics; (c) obliquity of magmatic state subfabrics.

The third of these criteria was developed further, mainly experimentally by a French group, and culminated in several papers on the work including Fernandez & Laporte (1991) and Ildefonse *et al.* (1992), which outlined shear sense criteria that can be applied in both the field and laboratory (Fig. 1.6c) even to granitoids that had a low percentage of phenocrysts at the time of deformation (see

Ch. 4 for application of this technique). The method uses the obliquity commonly formed between either different minerals or different families of the same mineral due to their different rates of rotation whilst undergoing non-coaxial deformation.

1.3.5.b *Sense of shear from solid state fabrics*

These can be applied to both granitoids that have been deformed in the solid state (see section 1.3.2) and wall rocks, assuming that they have not melted beyond the RCMP. The concept of S-C fabrics now widely used in metamorphic rocks (Lister & Snoke 1984), was first applied to granitoids in 1979 by Berthé *et al.* For the purposes of this thesis, the author prefers the terminology summarised in Platt (1984). It uses the obliquity between two sets of planar foliations, S (schistosité) and C (cisaillement, or shear plane). The S-surfaces (Fig. 1.7) are defined by the preferred orientation of the long and intermediate axes of ellipsoidal deformed grains or aggregates, and tabular or prismatic mineral grains. Such shape preferred orientation fabrics are often thought to represent the XY plane of the finite strain ellipsoid (Ramsay & Graham 1970) and will, therefore, vary in orientation within a shear zone according to the strain, but they are thought to form at an angle of 45° or less to the shear plane. They curve asymptotically into a set of discretely spaced surfaces parallel to the boundaries of the shear zone, which form as a result of periodic variations in the amount of shear strain, and are themselves small-scale shear zones. The width and spacing of these surfaces may vary between that of individual grains and that of the main zone. They are labelled C in Fig. 1.7 after Ponce de Leon & Choukroune (1980). The sense of shear along the C surfaces is synthetic to that of the main shear zone. Many shear zones also contain secondary planar fabrics that develop at a later stage. One or more sets of spaced, irregular and discontinuous surfaces may form obliquely to S; these were called 'extensional crenulation cleavages' by Platt (1979) and Platt & Vissers (1980). Unlike C they are normally oblique to the main shear zone. Ponce de Leon & Choukroune (1980) have shown that they overprint both S and C and called them C'. Platt (1984) called them 'ecc', and they are labelled as both C' and 'ecc' in Fig. 1.7. They may occur as single, multiple or conjugate sets, with single and multiple sets being synthetic with the main shear zone (see Fig. 1.7 & Platt 1984 for full description).

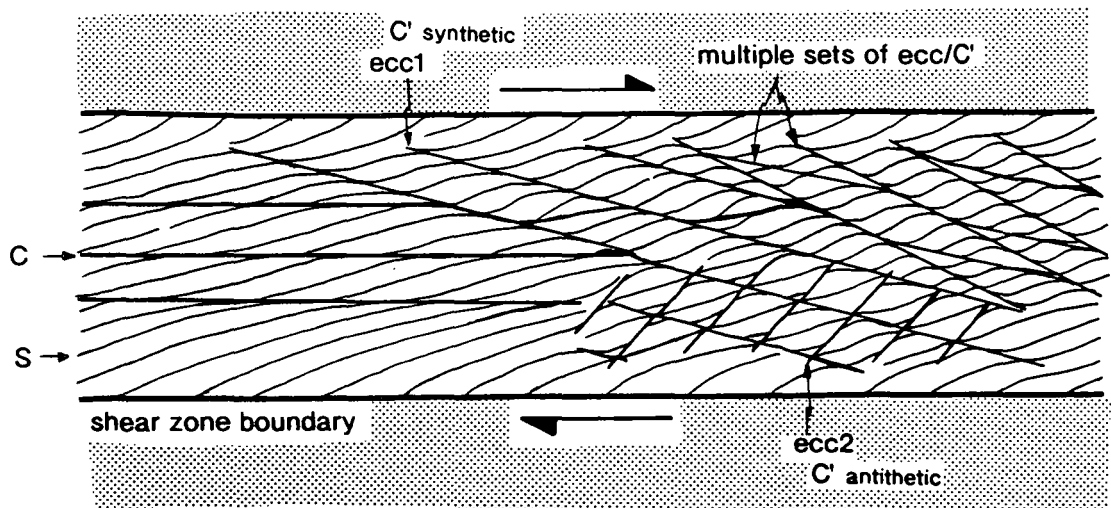


Figure 1.7 Shear band geometry as a method to deduce sense of shear in rocks deformed in the solid state (after Ponce de Leon & Choukroune 1980; Platt 1984).

Other shear sense indicators are summarised in Fig. 1.8 (after Simpson & Schmid 1983). These include asymmetric augen structures (σ or δ) and asymmetric pressure shadows on porphyroclasts due to the high ductility contrast between porphyroclast and matrix and rotation with the same sense of vorticity (Lister & Williams 1983) as the main shear zone. Curved inclusion trails in porphyroblasts growing in a deforming matrix can also be a useful shear sense indicator, furthermore, they can provide valuable time constraints on the relationship between metamorphism and wall rock deformation (Simpson & Schmid 1983; Bell & Johnson 1989). Crystallographic fabrics (Lister & Hobbs 1979; Simpson 1986; Law 1990) may also be used to determine shear sense. Quartz is the most commonly studied mineral in this type of work and scientists such as Law *et al.* (1986) and Gapais & Cobbold (1987) illustrate how quartz micro-fabrics can help differentiate between flattening and non-coaxial regimes and can also provide an estimate of the shape of the finite strain ellipsoid. Feldspar fabrics have occasionally been used in this kind of study, however, the biaxial nature of minerals such as plagioclase makes interpretation more complicated (Tullis & Yund 1985; Bell & Johnson 1989). Biotite and other phyllosilicates can provide some kinematic information, but in general their behaviour is poorly understood and hence they are rarely studied.

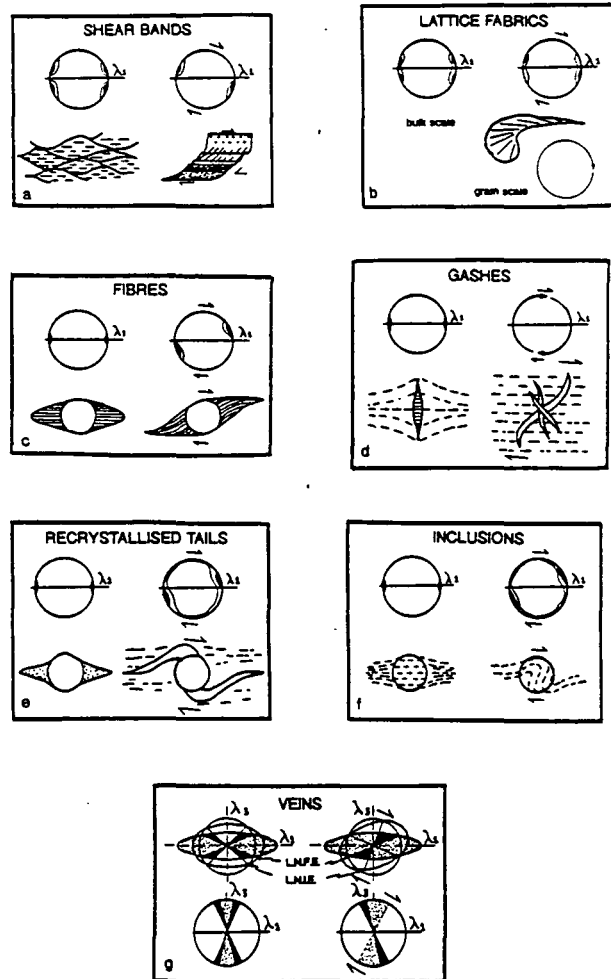


Figure 1.8 Summary sketch of solid state shear sense indicators (after Blumenfeld & Bouchez 1988). Sense of shear criteria relate to the orientation of (a) shear bands; (b) lattice fabrics; (c) fibres; (d) gashes; (e) recrystallised tails around porphyroclasts; (f) inclusion trails within porphyroblasts; (g) vein arrays.

The above shear sense criteria are mainly the product of ductile deformation. As mentioned previously, granitoids may exhibit deformation features which straddle the brittle-ductile transition, for example brittle features in feldspar coexisting with ductile features in quartz. In such cases displaced or broken grains may be shear sense indicators, as may kinking of twins and the orientation of submagmatic microfractures (Bouchez *et al.* 1992). However, often the shear on microfractures may be antithetic to the overall shear regime, so caution must be exercised. In some cases marker horizons may be displaced, indicating the sense and amount of deformation. Other brittle shear sense indicators include slickenfibres, striations and a variety of secondary fractures (summarised in Ramsay & Huber 1987; Petit 1987).

One must always exercise extreme caution not to apply only one shear sense criterion in isolation, since it is common to find zones of antithetic shear on various scales. It is wise to collect numerous data before reaching definite conclusions about the kinematics of deformation.

1.3.6 *The validity of the Cloosian approach*

The work of Hans Cloos (Balk 1937) saw the initiation of the study of fabric and structural analysis in igneous rocks. He subdivided granitic structures into primary and secondary structures. *Primary structures* were considered to have formed during magma consolidation as a result of igneous flow parallel to the wall rocks. This resulted in the formation of preferred orientations of minerals in planar or linear flow fabrics. Cloos also regarded joints as primary structures since they sometimes bear a geometrical relationship to the flow fabrics, hence they were regarded as a continuation of the magma deformation.

Secondary structures were considered to have formed during sub-solidus deformation and were thought to be essentially tectonic and regional in origin. Examples of secondary structures were believed to be foliations which cross-cut internal contacts and boudins.

The philosophy of primary igneous flow has been followed, even fairly recently (Marre 1986), but in the light of interpreting granitic structures in relation to regional tectonics much of this work has fallen from favour. It was first criticised by Berger & Pitcher (1970), then Pitcher & Berger (1972) and Hutton (1988a). These and other authors have been mainly concerned with the ability of granitic magma to flow in the conventional Newtonian manner suggested by Cloos and Balk, who regarded magma flow in much the same way as water flows. However, since silicate melts are considered to act as viscous materials (c.f. section on rheology) and can be treated as 'normal' geological materials, then it is difficult to comprehend how Cloos' primary structures could have formed in the manner suggested. It is possible, however, that such features may form due to the development of localised stresses around the wall rock - magma interface, but even in this case, the response of the magma would be that of a complex material and not a pure fluid as Cloos invoked.

In addition to the complex rheology of magma, Berger & Pitcher (1970) questioned the use of joints to constrain granite tectonic models, pointing to the difficulty in distinguishing primary jointing, formed during consolidation, and secondary jointing, which is often ascribed to the regional

stress field and uplift. They accepted that some joints may form early in the consolidation history, as is clearly evident from the presence of dykes and veins along joint systems within many plutons. However, they pointed out that 'primary' joints that do not contain dyke material are likely to be obliterated by late and post-consolidation processes which invariably alter the early igneous textures.

1.4 Emplacement Mechanisms

The field examination of a pluton aims to use all observable information on fabric development and kinematic indicators in order to develop an emplacement model. In order to produce a valid emplacement model one must take into account the structural and metamorphic development of the wall rocks as well as the granitic rocks.

Of particular importance is the timing of regional or local deformation and metamorphic events relative to intrusive events. Pitcher & Read (1959) and Pitcher & Berger (1972) were among the first to establish the timing of granite intrusion using the deformation history of the thermal aureole. This work was modified by Hutton (1977) and Paterson *et al.* (1989) in Donegal and the Sierra Nevada respectively. Modern kinematic analysis and dating techniques are useful tools to aid the establishment of the relative timing of pluton emplacement and deformation, whilst isotopic studies may be used to date events in the aureole. Page & Bell (1985) add a hint of caution to this type of study, reinforcing that one must take account of the possibility of the resetting of ages due to later deformation and metamorphism.

Structural maps produced from detailed field studies containing information about both igneous and wall rocks are often used to interpret regional structures and intrusion events with reference to pluton emplacement, and hence to produce an emplacement model. This approach could result in the production of only a local emplacement model (Paterson & Fowler 1993; also see Chapter 6) which may have little bearing on the emplacement of large plutons or batholiths, but such models may be of use in constraining larger scale models. Emplacement mechanisms can be very influential in controlling the shape of plutons on the ground. In the past there has also been a tendency to assume an ascent mechanism from emplacement information and whilst this may be valid in some cases, it is not always necessarily correct (Pitcher 1979).

Pluton emplacement is traditionally regarded in terms of whether it has occurred passively or forcefully.

1.4.1 Forceful emplacement

This involves deformation of the crustal rocks to create space for the magma. It may be due to diapiric ascent of magma, or may be a secondary feature associated with ballooning of plutons at the emplacement level. In both cases, there is evidence for expansion, with intense flattening strains developed in the pluton and its wall rocks. Examples of forcefully intruded plutons include Ardara (Pitcher & Berger 1972, Holder 1979), Papoose Flat (Sylvester *et al.* 1978, Paterson 1991), Cannibal Creek (Bateman 1984) and the North Arran granite (England 1988).

1.4.2 Passive emplacement

This occurs when crustal rocks are simply 'replaced' by magmas, often by stoping of large blocks, usually along pre-existing anisotropies, which physically creates space for magma. This mechanism was invoked by Daly (1933) to explain the occurrence of large discordant cauldrons in the Sierra Nevada batholith. Pitcher & Berger (1972) also employed a passive stoping model for two of the Donegal plutons, Thorr (the main study area in this project) and Fanad. Whilst it is clear that this is a feasible model, even on quite large scales, it may not always be a very efficient method of space creation since it requires that the stoped country rocks have a higher density than the magma. Furthermore, it results in fairly rapid heat loss from the magma to the wall rocks by conduction and therefore relatively rapid cooling and arrested ascent of the magma (Pitcher 1987).

1.4.3 Tectonic controls on emplacement

The traditional approach outlined above fails to account for the increasingly recognised association between magma emplacement and active tectonics. In cases where this interaction is rather 'passive', active tectonism results in the creation of a space into which magma will then be emplaced. An example of this is Strontian (Hutton 1988b) where an extensional shear zone created a flat which granite then filled. The formation of ring dykes, cauldrons and cone sheets are other

examples of passive emplacement where the magma stresses interact with the earth's free surface (Clough 1909; Anderson 1936).

The most recent work suggests that magma ascent and emplacement may be more intimately and actively associated with tectonic processes than is implied by all of these earlier models. In particular, sheeted structures have been recognised increasingly in fault and shear zones (D'Lemos *et al.* 1992; Hutton 1992; Hutton & Reavy 1992). In addition to the examples given by these authors, other sheeted plutons include the Mortagne pluton (Guineberteau *et al.* 1987) emplaced into a dilational zone associated with sinistral shear in the South Armorican shear zone and the NE Gander zone granites (Holdsworth 1991, Holdsworth *et al.* 1993) which are mostly associated with sinistral shear along the Dover Fault zone. It is now common and useful to subdivide granitoids emplaced into active shear zones on the basis of whether the kinematics of intrusion were extensional, compressional or strike-slip dominated. The examples cited above are plutons emplaced into strike-slip regimes. One of the best known examples of emplacement in an extensional shear zone is the Strontian granite (Hutton 1988a & b). Plutons emplaced into compressional shear zones have been recognised by Blumenfeld and Bouchez (1988). These plutons are associated with thrust zones formed during the Hercynian orogeny. Another example of pluton emplacement in such a setting is the Great Tonalite Sill, Alaska (Ingram 1992).

In general magma emplacement style in tectonically active zones is controlled by the interplay of extensional strain rate and magma buoyancy rate (Hutton 1988a), such that passive emplacement can be related to a surplus of extensional strain rate resulting in dilation so that magma is effectively 'sucked' into the resulting cavity; forceful mechanisms, on the other hand, can be related to a deficit in the extensional strain rate relative to the magma buoyancy rate.

1.5 Aims, objectives and structure of the thesis

The most basic objective of this thesis is to add to our understanding of the interplay between active tectonism and magma emplacement. Part II (Chs. 2,3 & 4) of the thesis is a detailed study of the Thorr granite, County Donegal. In addition two other plutons (Ratagain, Inverness-shire; Loch Loyal, Sutherland) were examined in shorter reconnaissance studies; these are outlined in Part III of

the thesis (Chs. 5 & 6). New emplacement models are proposed for the three plutons. The emplacement models are compared in order to determine whether there are any common factors that have played a role in the emplacement of all three plutons, and hence may be important in relation to granite emplacement in general. In addition, the nature and evolution of deformation fabrics within each of the plutons is outlined in order to assess the relative importance of deformation related to emplacement, compared to that produced by the overall tectonic regime.

A new criterion for determining shear sense in rocks deformed in the magmatic state is outlined in the thesis. A further aim of this thesis is to assess the usefulness of this technique in the field.

1.6 An introduction to the Donegal Batholith

The Donegal batholith is a classical area for studying granite emplacement and it has been the topic for considerable debate mainly due to contention over whether its origin was igneous or metamorphic. Pitcher and Berger (1972) finally disproved the granite series approach supported by Read and Buddington by showing that the plutons of the Donegal batholith, although at the same crustal level, were emplaced in significantly different ways. Hutton (1982a) developed this idea and produced a model which advocated emplacement of all seven plutons in relation to the sinistral shear zone into which the Main Donegal granite was emplaced. The model relied on the fact that an active shear zone can produce areas of extension and compression in bend and tip zones (Sanderson & Marchini 1984) and that the Donegal plutons could be related to such areas along the main shear zone. It follows that dilational zones produce large passively emplaced plutons containing many stoped blocks, examples include Thorr and Fanad. Compressional zones then produce large forcefully emplaced plutons such as Ardara and Toories. The Main Donegal granite exhibits a characteristic sheeted form due to wedging of magma into the shear zone.

Part II

***THE EMPLACEMENT AND DEFORMATION OF THE THORR
GRANITE***

Chapter 2

THE THORR GRANITE AND ITS COUNTRY ROCKS

Introduction

Part two of this thesis is concerned with the emplacement of the Thorr granite; thought to be the earliest of the plutons in NW Donegal, which together form the Donegal batholith (see Fig. 2.1). Before considering the emplacement events it is necessary to relate them to the regional geology in terms of chrono- and tectonostratigraphy, as defined by many investigators prior to this work.

2.1 The Caledonian orogeny in NW Donegal

NW Donegal lies roughly half way along the trace of the linear Caledonian orogenic belt of the North Atlantic region, which can be traced from Svalbard and NE Greenland, down through the British Isles and Newfoundland, into Maritime Canada and the eastern USA (Harris *et al.* 1979). This orogen is thought to have developed during the evolution and destruction of a Lower Palaeozoic ocean, Iapetus (Dewey 1969). Donegal lies near to the margin of the largest continental landmass, Laurentia (which comprises much of North America, Northern Britain and Greenland). Deformation associated with final closure, mainly during the Silurian according to the most recent synthesis (Soper *et al.* 1992), has segmented the orogen in Britain into a number of fault-bounded terranes. The Donegal region forms a large part of the Grampian Terrane in NW Ireland, bounded to the NW by the Great Glen Fault and to the SE by the Highland Boundary Fault - Clew Bay Line (Hutton 1987). The terrane comprises Late Precambrian metamorphic rocks of the Dalradian Supergroup that are intruded by a series of Siluro-Devonian Granites, forming the Donegal Batholith (Pitcher & Berger 1972).

The granites post-date the main regional phases of folding in the Dalradian country rock, but their intrusion overlaps steep sinistral shear zones formed during terrane displacements late in the orogenic cycle (Hutton 1982; Soper & Hutton 1984). The ages of the earlier deformation and

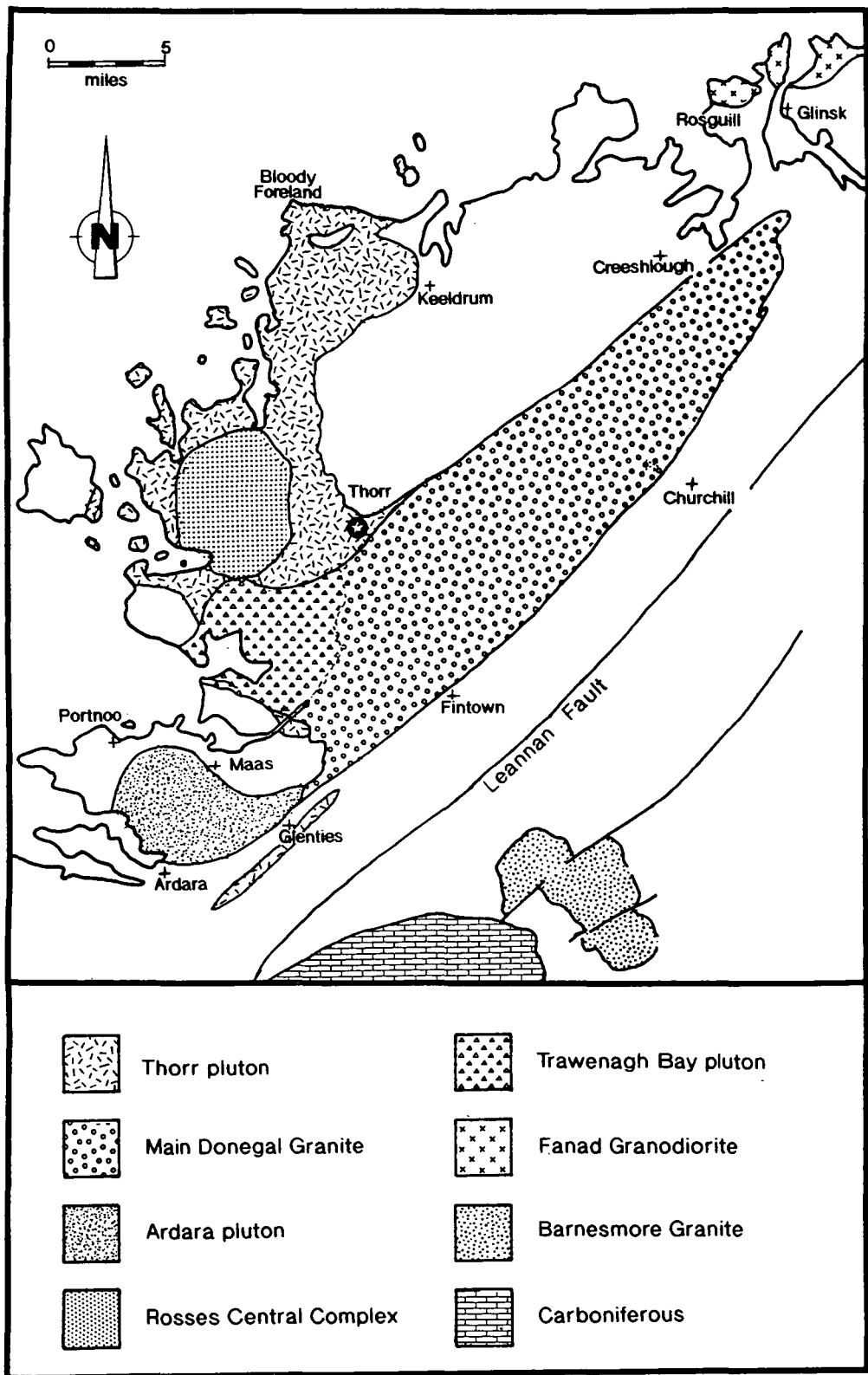


Figure 2.1 Map of NW Donegal showing the area of the Donegal Batholith. The main plutons are referred to in the key; the unshaded area around the batholith indicates the occurrence of Precambrian (Dalradian) country rocks.

metamorphism in the Dalradian are problematic, especially in Scotland (e.g. see Rogers *et al.* 1989 for discussion), but in Ireland, regional D_3 deformation (see section 2.4) *must* be Caledonian because this event is associated with SE-directed thrusting of Dalradian rocks over the Ordovician volcanics of the Tyrone Igneous Complex in Tyrone along the Omagh Thrust (Alsop & Hutton 1993).

2.2 Previous work in NW & Central Donegal

There is a wealth of literature about NW Ireland which dates as far back as 1834 with the work of M^c Adam. Even at an early stage, the geology of NW Donegal, particularly the origin of the granites, was a highly controversial issue. The debate centred around the fact that many workers failed to recognise that the batholith was composed of many plutons. Around the middle of the last century a metamorphic origin was favoured by many who studied the batholith (e.g. Scott 1862; Blake 1862; Harte 1867 and Green 1871). A purely metamorphic origin was proposed as late as 1881 by Hill, who believed that the granite was a Laurentian gneiss which was unconformably overlain by a metamorphosed sedimentary sequence. Haughton (1862) proposed a combination of metamorphic and igneous origin. It was only in the latter part of the last century that an entirely igneous origin was proposed for the granite, advocates of this were Calloway (1885) and Hull *et al.* (1891). In the memoir of the Irish Geological Survey Hull stated his belief that

"there were at least four or five periods of granite intrusion."

Cole (1902, 1905, 1906, 1916) was the only person to work on the granites of Donegal in the early part of this century. Between 1916 and 1948 there was a lull in activity, until 1948 when H.H. Read established a research team to examine the granites and their country rocks. This generated a large database during a period of about twenty years. The main contributors were Pitcher (1953a, 1953b); Gindy (1953); Iyengar *et al.* (1954); M^c Call (1954); Whitten (1951, 1955, 1957a, 1957b, 1960, 1966); Akaad (1956a, 1956b); Pitcher & Sinha (1958); Pitcher & Read (1959, 1960, 1963); Mercy (1960); Knill & Knill (1961); Rickard (1962); Pitcher & Shackleton (1966). In addition, there have been many PhD. theses, a full list of which are listed in Pitcher & Spencer (1968). The main findings of the theses and

various other papers, including Berger (1967), Naggar (1968), Edmunds (1969), Fernandes-Davila (1969), Naggar & Atherton (1970) and Wood (1970) were compiled in Pitcher & Berger (1972).

Work in the area since 1972 has been sporadic but varied, with investigation of geophysical aspects (Young 1974; Evans & Whittington 1976), geochronology (Fitch & Miller 1980; O'Connor *et al.* 1982, 1987) and granite emplacement (Holder 1979; Hutton 1982). Recent relevant theses include Scott (1974), Atkin (1977) and Oglethorpe (1987). Oglethorpe investigated the detailed geochemistry and petrology of the Thorr granite and his work will be referred to in this thesis where relevant, especially with reference to petrology and geochemistry.

2.3 Stratigraphy in NW Donegal

The stratigraphy of the Dalradian is thought to have recorded the formation and progressive foundering of an ensialic to passive margin basin located on the SE boundary of the Laurentian continent during the late Precambrian (Anderton 1985). Early shelf dominated sediments (Appin - Argyll Groups) give way progressively to a deeper water turbiditic facies (S. Highlands Group) with increasing tectonic instability and intrusion of widespread tholeiitic sills, reflecting increasing amounts of rifting and crustal extension through time (Harris *et al.* 1978; Anderton 1985).

The Thorr pluton is emplaced into Dalradian metasediments, specifically, the upper formations of the Creeshlough Succession from the Ards quartzite to the Upper Falcarragh pelites that form part of the Lower Dalradian (Appin Group) succession in NW Donegal (Pitcher & Shackleton 1966; Harris & Pitcher 1975; see also Fig. 2.2). The Creeshlough Succession (Pitcher & Shackleton 1966) comprises nine formations which have a maximum thickness of 5000 metres, although the observed thickness is often very variable due to a combination of tectonic thinning and ductile faulting (sliding) between stratigraphic horizons of different competencies and complex folding and repetition of other units.

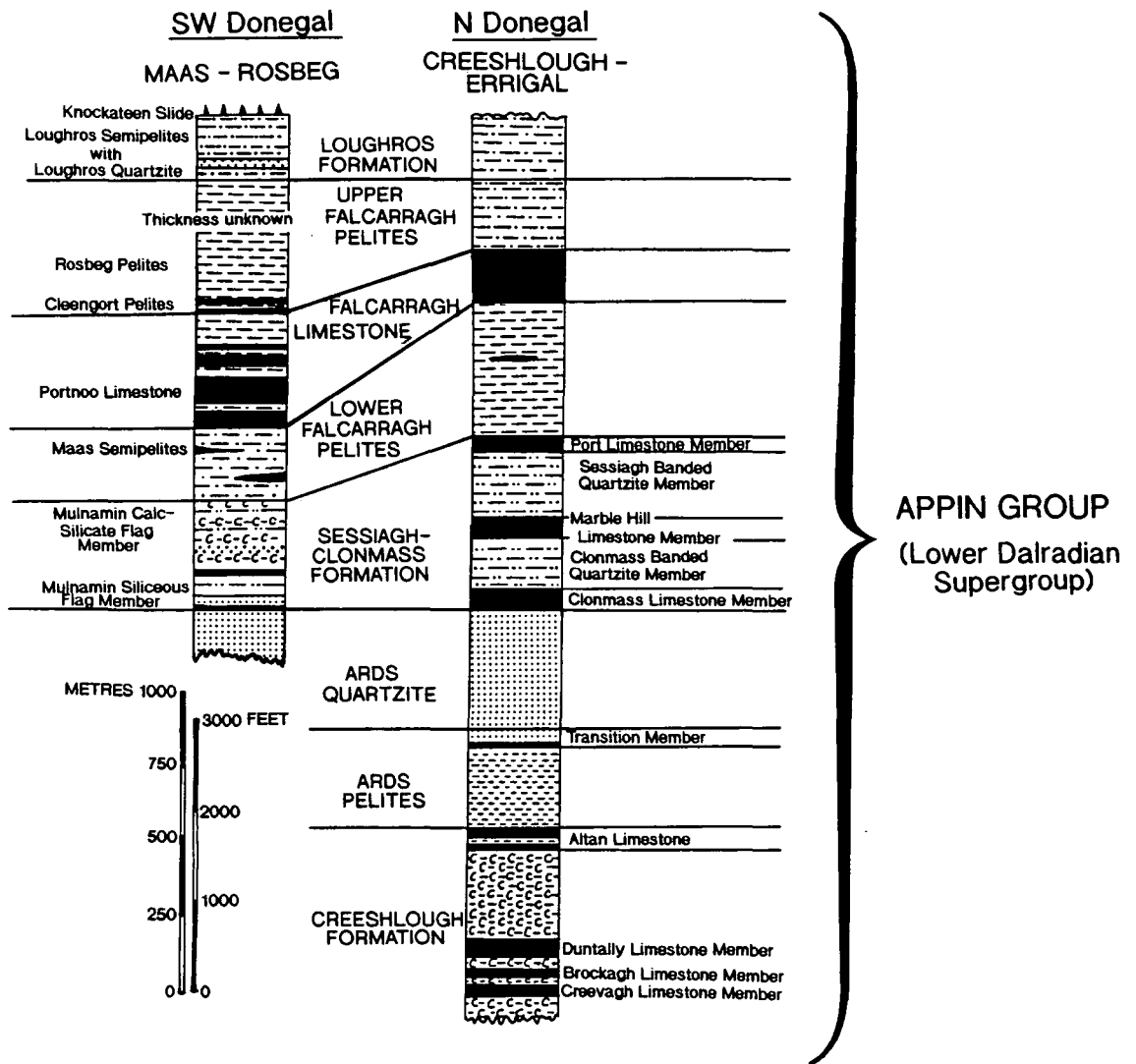


Figure 2.2 The Creeshlough Succession in North and Southwest Donegal (after Pitcher & Berger 1972).

The sediments of the Creeshlough Succession and the overlying Kilmacrennan Succession represent a gradual change in facies from a combined quartzite - black shale association upwards into a graded quartzite affiliation and then into a turbiditic facies. The lowest horizons generally consist of well-differentiated and well-sorted formations, with characteristic coarse-grained, clean, feldspathic sandstones. Individual beds may be continuous for kilometres without interruption. Sedimentary structures include large-scale cross-lamination, very coarse grading in pebbly bands and mud cracks, suggesting that some units were deposited under near-surface conditions. Furthermore, there is evidence for winnowing at the top of individual beds of tillite and local shallows. The conclusion is then, that these are relatively shallow water deposits (Pitcher & Berger 1972). This includes the black shales, since these are rhythmically interbanded with sandy beds over a considerable thickness; a particularly good example of this is seen at the base of the most typically shallow water deposit of all, the Ards quartzite, one of the host rocks of the Thorr pluton. The occurrence of pure limestones and dolomites, which may alternate with quartzite flags or be interbedded with tillites, confirms the claim that the successions are representative of relatively shallow water successions (Pitcher & Berger 1972).

Pitcher and Berger (*op. cit.*) also state that there are significant variations in facies thickness which are consistent with the concept of "differential subsidence" involving the presence of troughs and basins where sedimentation was more continuous than on the flanks and walls; the latter were often quite near to the surface. Anderton (1985) interpreted such changes in thickness in the Dalradian (mainly in Scotland) as being due to the evolving tectonic architecture in the sedimentary basin. In essence, thickness variations can be interpreted by the process of differential subsidence and uplift of small fault blocks defined by Caledonoid-trending listric normal faults and trans-Caledonoid transfer faults (Anderton 1985).

There is little or no evidence as to the direction of flow during sedimentation, except at Rosguill, where primary structures indicate sediment transport from the west and north (Knill & Knill 1961, p.277).

Substantial magmatism is contemporaneous with and succeeded Dalradian sedimentation (Pitcher & Berger 1972). Pillow lavas are associated with the turbidites, however extrusives are not

found. An extremely large volume of tholeiitic magma is represented by numerous sills of varying dimensions, which were emplaced shortly after sedimentation but before the majority of the deformation (Pitcher & Berger 1972). The distribution of the most prominent bodies is illustrated in Fig. 2.3.

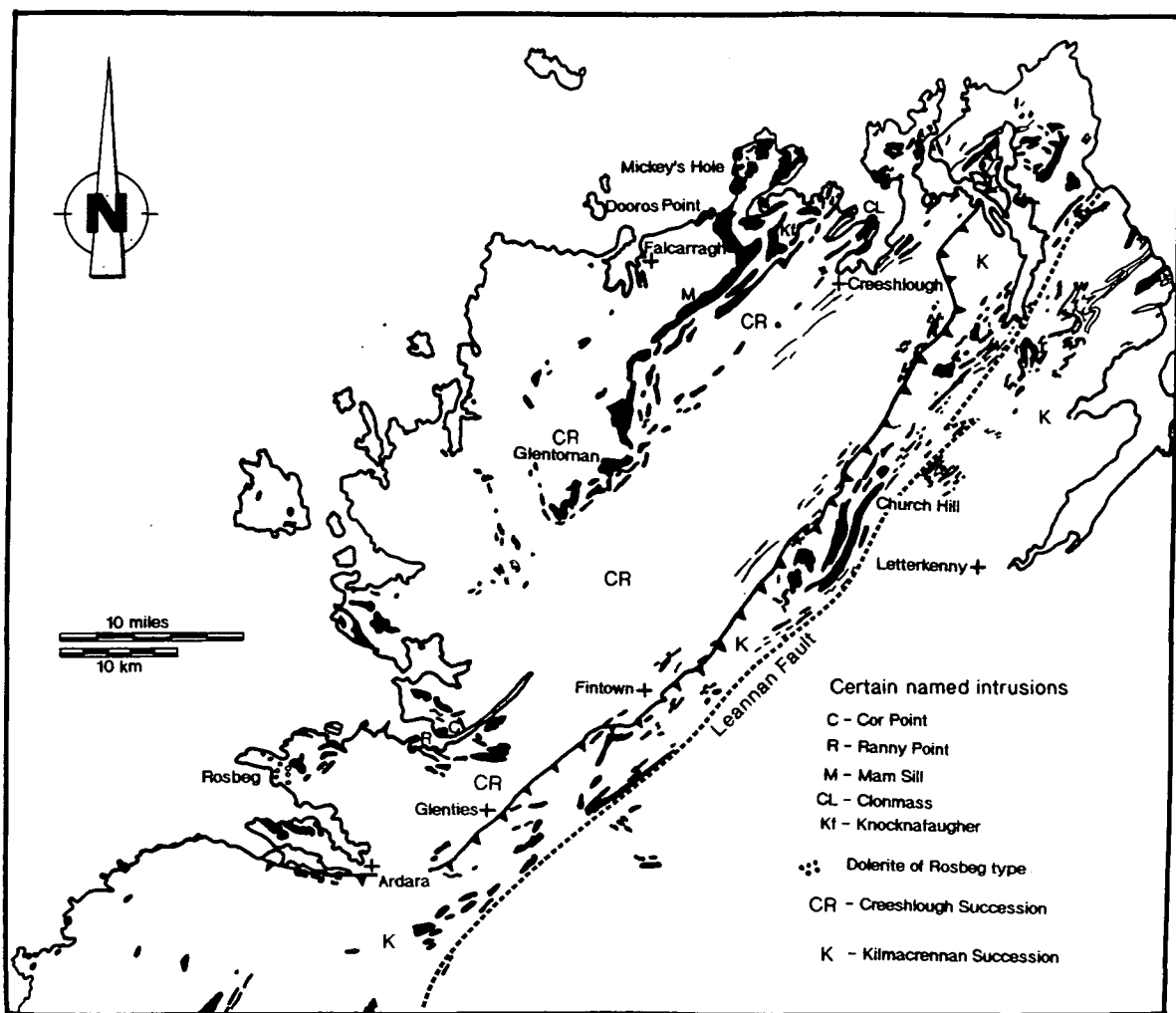


Figure 2.3 Map showing the occurrence of the largest Dalradian age basic sills (now metadolerites) in Donegal (after Pitcher & Berger 1972).

The volume of magma emplaced in Donegal exceeds the volume emplaced anywhere else in the Dalradian, with the possible exception of the Argyll coast. The sills are mainly quartz dolerites, which are now metadolerites due to regional and limited thermal metamorphism; these have simple contact aureoles. However, a slightly different type of sill is found around Rosbeg. Again these are also doleritic, but their present chemistry and mineralogy is slightly different, being represented by very coarse-grained garnet amphibolites, often with a distinctive thin skin of white adinole. With reference to the precise time of intrusion relative to sedimentation and deformation, Obaid (1967) showed that the Rosbeg type dolerites predate the earliest phase of deformation (see section 2.4). However, the quartz dolerite sills were often emplaced between the first and second deformation event, as illustrated by the Maam sill, which cuts the Horn Head slide (Pitcher & Berger 1972; Hutton 1979). The present differences between the two types of dolerite, consequentially, can be attributed to the contrast in their metamorphic history, since both are tholeiites in terms of their mineralogy and chemistry.

2.4 Structures in the Creeshlough Succession

This involves a complicated, polyphase deformation history during the Caledonian orogeny. The exact chronology of deformation has been the subject of much debate, particularly during the past forty years. Table 2.1 summarises most of the key works from this period concerning the structural evolution of NW Donegal. The major problem is the correlation of structures across an area of approximately 80 km², in which there are quite obvious highly localised deformation phases, some of which are related to intrusion of specific granite plutons.

As Table 2.1 shows, the first deformation event involves the development of a pervasive 'bedding-schistosity', or S_1 cleavage. Locally, however, the bedding-schistosity can be seen to cross-cut bedding without crenulating it (Hutton 1977, 1979). It is unclear, however, whether D_1 is related to any larger-scale structures. Knill & Knill (1961) and Pitcher & Berger (1972) argued that it is related to the development of major bedding-parallel detachments, or tectonic slides. Others, such as Rickard (1962) and Hutton (1977, 1979, 1982a), argue that these slides are not D_1 in age, but D_2 . Hutton (1979) reports

KNILL & KNILL (1961)	HOWARTH et al. (1966)	CAMBRAV (1969a)	PITCHER & BERGER (1972)	RICKARD (1962) & (1963)	HUTTON (1979) (1982a)	HUTTON (1977)
S2* = Bedding-schistosity + slides "major fold of Rossguill" (?)	F1	F1	D1, S1	Bedding-schistosity	D1, S1, F1 (?)	D1, S1, F1 (?)
Errigal syncline with S3 (+ S3(?)) cleavage	F2	F2	D2-D3 Composite structures ┌ F2, S2 └ S3	Major folds (L1), slides and crenulation cleavage Crockator cross-fold (?)	D2, S2, F2	D2, S2, F2
Flat S4 cleavage				Dunlewy cross-fold (?)	D3, S3, F3	D3, S3, F3
oblique cleavage S5(?)	F4(?)	F3 F4	D4 ┌ S4a └ F4b	Bending of Crockator cross-fold around Thorr Bend. Related to Thorr pluton Cashel belt of superimposed folds (L2) due to Thorr granite	D4, S4, F4 Upright, N-S trending folds and cleavages related to Thorr & Ardara D5, S5, F5	D4, S4, F4
		F5(?) late angular folds	DEFORMATION ASSOCIATED WITH THE MAIN DONEGAL GRANITE DMG1 mineral alignment & mullions DMG2, SMG2, FMG2	Devlin belt of superimposed folds (L2') due to Main Donegal Granite	D6, S6, F6 Steep NE-SW cleavages & upright folds; highest strain; peak thermal metamorphism	
			?			
			DMG3, SMG3	Later fold phases F4-F8 as yet uncorrelated with late events in other areas	D7, S7, F7 = Steep NE-SW cleavages & Upright folds D8, S8, F8 } Conjugate D9, S9, F9 } cleavages & kink bends	D6, S6 D7, S7, F7

* S1 = Bedding

Table 2.1 Summary table of published deformation chronologies.

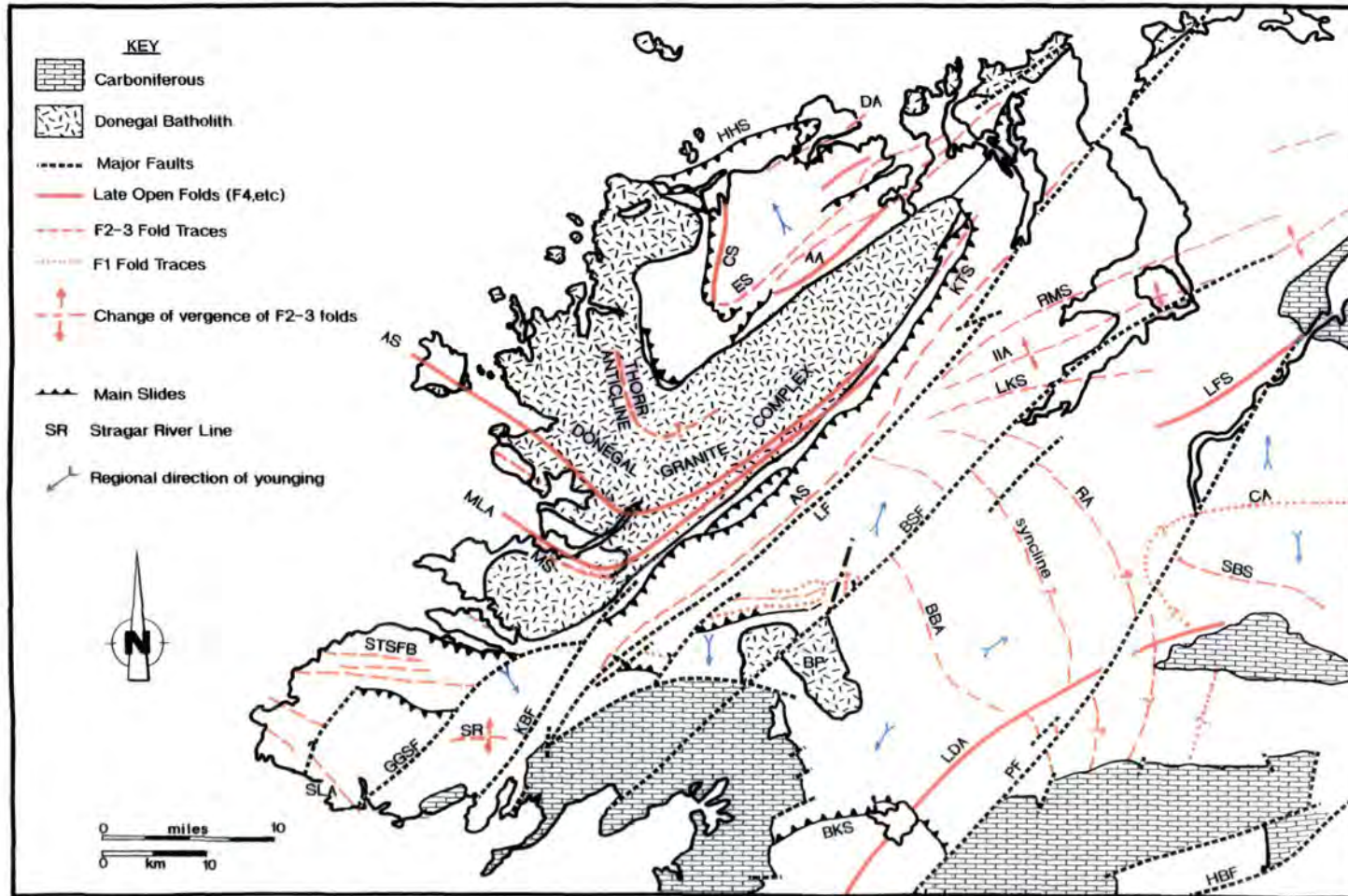


Figure 2.4 Map illustrating the occurrence of some pre-granite intrusion deformation structure as summarised by Pitcher & Berger (1972).

that no minor F_1 folds have been observed, but that the vergence of the first cleavage on bedding is approximately towards the north, with bedding facing northwest in S_1 (although Hutton (1977) states that it faces south). Pitcher & Berger (1972) believe that some F_1 folds are recognisable around Ballybofey (see Fig. 2.4).

Pitcher & Berger (1972) argued that the next deformation event produced a set of composite structures, involving major recumbent folds (F_2) and an associated crenulation cleavage (S_2), followed by the development of another cleavage (S_3). The peak of metamorphism was placed in the interval between D_2 and D_3 . Rickard (1962) and Hutton (1977, 1979, 1982a), however, successfully separated Pitcher & Berger's composite structures. According to these authors, the D_2 event produced crenulation cleavages which are axial planar to occasional tight minor folds. Both the crenulation cleavage and the fold axial planes are inclined gently to steeply towards the south. Fold axes and bedding-cleavage intersections often trend E-W, although there is considerable variation. The plunge of minor F_2 fold axes is also very variable (Meneilly 1982). In general, the vergence of minor structures is towards the south and bedding faces north in S_2 (Hutton 1977). Examples of F_2 folds are given in Figure 2.4; they include the Errigal syncline (marked by the change in vergence of S_2 across the structure) (Rickard 1962; Pitcher & Berger 1972; Hutton 1977), the Crohy and Maghera (Meneilly 1982) and possibly Thorr anticlines; the Maghera and Thorr anticlines will be discussed further in Chapter 3.

In addition to these major folds and crenulation cleavages, Rickard (1963) suggested that the regional strike swing of the Dalradian rocks in NW Donegal, from the normal NE-SW Caledonian trend to the NW-SE trend seen around Thorr, Gweebarra and further south and west, may have initiated during the D_2 deformation event. The mechanism for the change in strike swing involved the development of cross-folds, such as the Crockator cross-fold, at Crockator, east of Thorr (see Fig. 2.5 & Map 1). The cross-folds were thought to be 'bent' later in the deformation history in relation to emplacement of the Thorr granite.

Rickard (1962) and Hutton (1977) recognised discrete D_3 structures, which cross-cut the earlier folds and cleavages. S_3 may cross-cut bedding at up to 40° and is associated with open to close minor folds. Intersection lineations plunge moderately to the south and southwest, with S_3 and F_3 axial

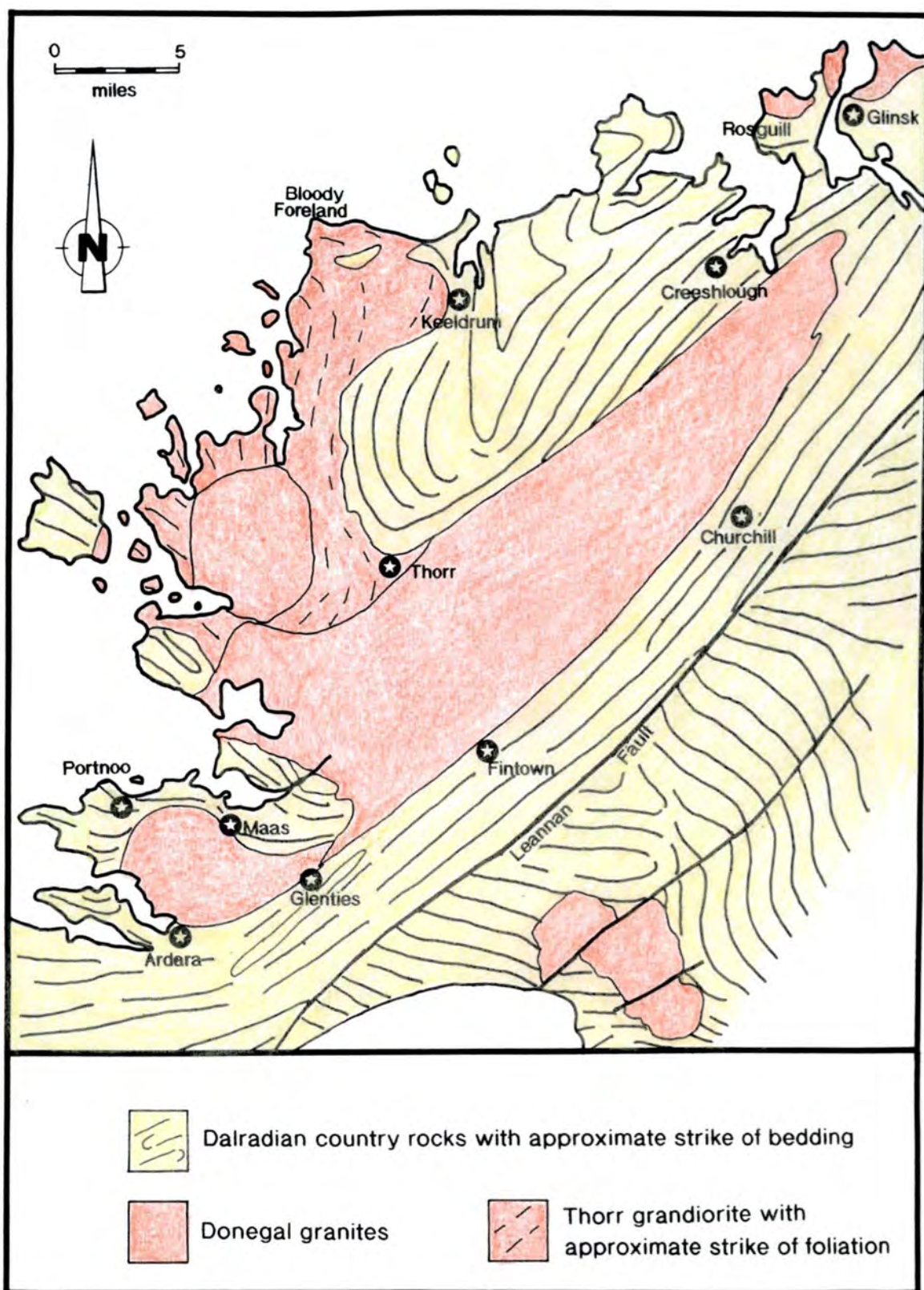


Figure 2.5 Sketch diagram illustrating the strike swing in the country rocks in Donegal (adapted after Pitcher & Berger 1972).

planes being subhorizontal to moderately inclined to the south. Vergence of D_3 structures on bedding is generally towards the east and southeast (Hutton 1977, 1979). These are quite clearly distinct from the D_2 structures, having opposite facing and vergence directions on the normal limbs of the F_2 folds. Rickard also placed the development of the Dunlewy cross-fold within the D_3 deformation event (see Fig. 2.6).

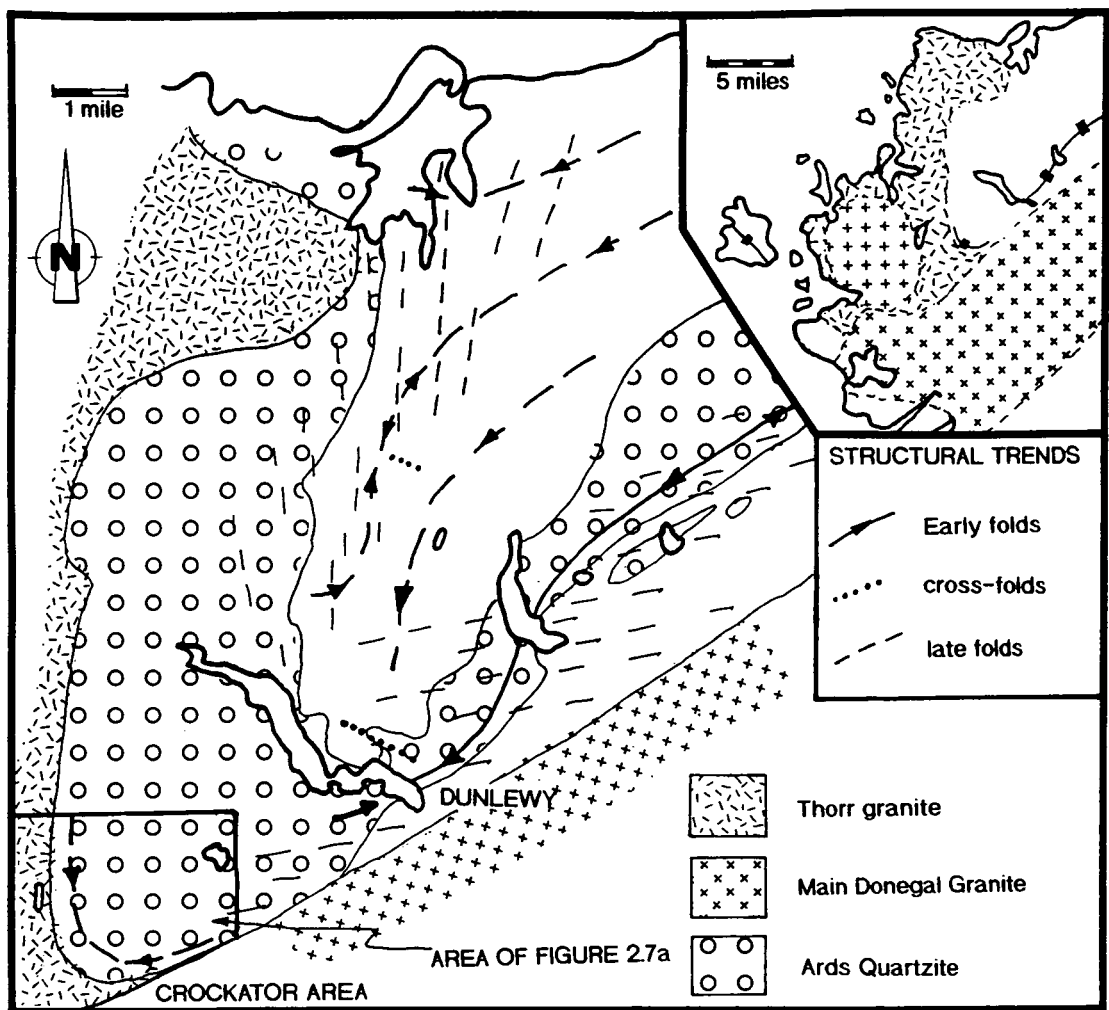
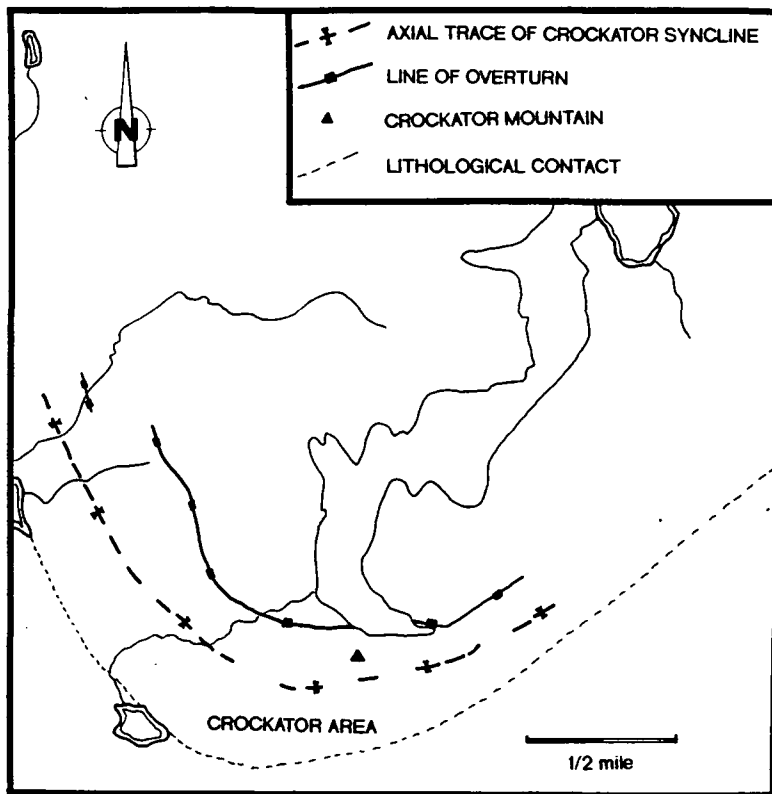


Figure 2.6 Sketch map of the area around Dunlewy and Crockator through to the north coast of Donegal showing the occurrence of early-, cross- and late- fold structures (after Rickard 1962). The figure illustrates the change in orientation of fold structures through time, which is reflected in the strike swing across the area, and is particularly well-illustrated at Crockator.

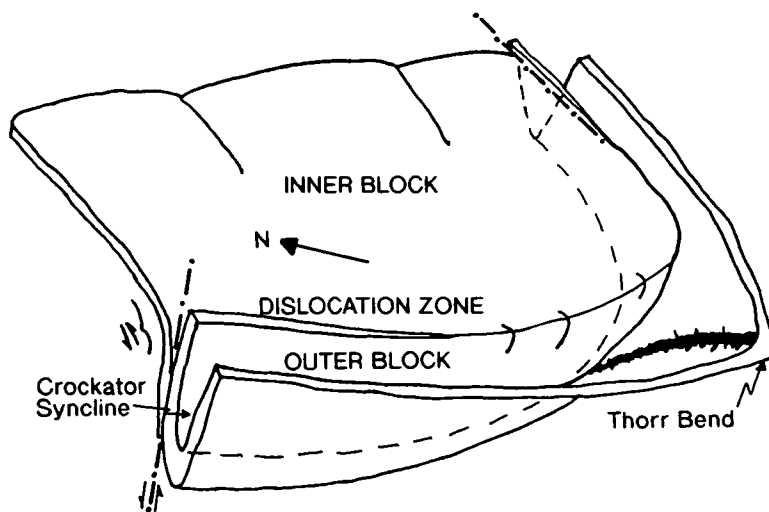
It is at this point that the correlation of structures across the whole region becomes problematic, since it was around this time that granite intrusion began, creating localised deformation events. Table 2.1 is an attempt to illustrate the complexity of the problem. For example, Hutton (1977, 1979) described a late, steeply inclined crenulation cleavage, which is well developed in the Creeshlough area within the aureole of the Main Donegal Granite, which he designated as being a D_4 structure. However, the same author (1982a) stated that although locally a D_4 cleavage, in the context of Donegal as a whole, it is actually D_6 in age. Rickard (1962, 1963) and Hutton (1982a) suggested that localised D_4 and D_5 structures may be related to the intrusion of the Thorr and Ardara plutons respectively. Both authors described these structures as upright, N-S trending folds and cleavages, as seen, for example, in the Cashel belt of superimposed folds (Rickard 1962) (see Figs. 2.4 & 2.6). In addition, Rickard (1963) related the accentuation of the strike swing in NW Donegal to granite emplacement. For example, at Crockator, east of Thorr (Fig. 2.7a & b) where the swing is most acute, bending of the Crockator cross-fold around the Thorr bend, resulted in reorientation of all pre-existing structures towards a N-S trend. However, this bending clearly predates granite intrusion, as seen by the N-S trend of the ghost stratigraphy within the Thorr pluton (for further discussion see Ch.3).

All later deformation phases have been related to the Main Donegal Granite Shear Zone (MDGSZ) by Hutton (1979, 1982a), who cross-references his chronology of later deformation events with that of previous authors (see Table 2.1 for summary). Using his chronology, D_6 represents the peak of deformation associated with the shear zone and with synchronous granite intrusion; it also corresponds with the peak of thermal metamorphism in the area. D_7 produces infrequent minor folds, restricted mainly to the Creeshlough area where they are D_5 in age (Hutton 1977, p.102). D_8 is the only episode of the very late deformation events which is of comparable development and importance to the major shear zone deformation event (D_6). It produces a steeply inclined, NNE-SSW trending crenulation cleavage of 'normal' type (Cosgrove 1976) and sinistral offset. This is often associated with a less frequent conjugate normal crenulation cleavage, trending WNW-ESE, and having dextral offsets (Hutton 1982a, p.620). These are equivalent to the S_6 cleavages of Hutton (1977, p.102) and the cross-cleavages of Pitcher & Berger (1972, p.260) and their geometries appear to be consistent with that of sinistral shear bands.

Figure 2.7 The geology of the Crockator area (after Rickard 1963).



(a) Map showing the axial trace of the Crockator syncline, including the trace of the line of overturn (after Rickard 1963).



(b) Three dimensional shape of the Crockator syncline (a cross-fold) which is 'bent' around the Thorr bend (after Rickard 1963). Rickard believed that it was bending of this fold that resulted in the pronounced strike swing in the area around Crockator.

They are restricted to, and are, therefore, assumed to be related to the shear zone. The development of such 'normal' or 'extensional' cleavages is suggested to be typical of late deformation in shear zones (Platt & Vissers 1980). Finally, D_9 is associated with sporadically developed conjugate 'reverse' kink bands (Cosgrove 1976), which are equivalent to D_7 (Hutton 1977) in the aureole of the granite, near Creeshlough.

The correlation of deformation events across the whole of Donegal, although problematic, is crucial to the study of granite emplacement events, since it is essential to decide what pre-dates and post-dates the emplacement of individual plutons. It is not an aim of this thesis to determine an exact deformation chronology on a regional scale. Instead, the present author has used key reference points within the existing chronology to date the deformation structures associated with the emplacement of the Thorr pluton. These reference points arise mainly from the work of Hutton (1982a) and Rickard (1962, 1963), who developed broadly similar deformation chronologies. Further details will be discussed in more detail in Chapter 3, but in the present context, the emplacement of the Thorr granite is considered to be post- D_3 and pre- D_6 .

In addition, the present author will avoid the use of D -numbers in later chapters, in favour of a more general terminology, since the occurrence of localised progressive deformation events in some areas, but not in others, can render the use of D -numbers both difficult and confusing.

2.5 Metamorphic history of the Creeshlough Succession

The rocks of the Creeshlough succession have experienced a combination of regional, low-grade metamorphism, associated with the earlier deformation events described in the previous section, and thermal, higher grade metamorphism, associated with the emplacement of the granitic magmas of the Donegal batholith. It is convenient to describe the regional and thermal events separately.

2.5.1 Regional metamorphism of the Creeshlough succession

As stated previously, this is a fairly low-grade event, with the peak of metamorphism being no greater than greenschist (Hutton 1982a) to lower amphibolite facies (Pitcher & Berger 1972). Pitcher & Berger (1972) attempted to correlate each of five regional metamorphic events with regional deformation events using the terminology of Sturt & Harris (1961); that is where MS_n refers to metamorphism synchronous with deformation event 'n' and MP_n refers to metamorphism post-dating deformation event 'n' (see Table 2.2). Hutton (1982a) attempted a cruder correlation, in terms only of the peak of regional and thermal metamorphism relative to deformation events (see Table 2.2).

The earliest event in the regional metamorphism involved the growth of minute phyllosilicates, quartz and feldspar along the S_1 cleavage plane; Pitcher & Berger (1972) called this MS_1 . Meneilly (1983) identified a similar metamorphic event, MS_1 in the Portnoo - Rosbeg area which again involves the development of muscovite and ores. The peak of regional metamorphism (MP_2 , Pitcher & Berger 1972) is syn- to late- D_2 according to Pitcher & Berger (1972), Hutton (1982a) and Meneilly (1983) and involves the development of garnet (now retrogressed to chlorite) and plagioclase porphyroblasts, along with occasional crystals of ilmenite, approximately contemporaneously with the D_3 event (Pitcher & Berger 1972; Meneilly 1983) (see Table 2.2). These porphyroblasts often contain inclusion trails of earlier cleavages, defined by trails of graphite dust particles and minute crystals of phyllosilicates, quartz and ores (Pitcher & Berger 1972).

The timing of regional metamorphic events after MP_2 is not easy to establish, partly because of the intrusion of the granite plutons, to which some of the metamorphism may be attributed. According to Pitcher & Berger (1972) the latest regional metamorphic events are fairly distinctive. Firstly, there was the regional development of random biotite porphyroblasts, with occasional growth of chlorite, which overprints D_2 and D_3 . Grain growth of quartz was also involved according to Edmunds (1969) and regional garnet was replaced by biotite at this time. The timing of this event relative to D_4 in the Creeshlough succession is unclear, but in the Slieve League area it clearly post-dates D_4 ; it has therefore been called the MP_4 event by Pitcher & Berger (1972). Across the whole area, regional garnet and biotite (both primary and after garnet) are replaced by chlorite to varying degrees during a retrogressive event

RMP_4 . According to Pitcher & Berger (1972) and the chronology outlined above, this event must succeed MP_4 , since it causes retrogression of the biotite flakes which grew during MP_4 . Within many of the granite aureoles, the chloritised garnet and biotite are overgrown by new thermal biotite, demonstrating that the regional retrogression, RMP_4 , occurred before the intrusion of the granites (Pitcher & Berger 1972). This is also the case with certain of the pyrite and carbonate porphyroblasts which may have formed during the retrogression, or which may represent an even later metamorphic event. However, the occurrence of new thermal amphibole after the regional carbonates in the aureoles of several of the appinitic complexes associated with the Donegal granites indicates that even the development of carbonate porphyroblasts predated granite intrusion (Pitcher & Berger 1972).

DEFORMATION (c.f. Table 2.1)		METAMORPHISM	
PITCHER & BERGER (1972)	HUTTON (1979) (1982a)	PITCHER & BERGER (1972)	HUTTON (1979) (1982a)
D1, S1	D1, S1, F1(?)	MS1 Growth of phyllosilicates along S1	Peak Regional
D2-D3 <div style="display: inline-block; vertical-align: middle;"> { <div style="display: inline-block; vertical-align: middle; margin-left: 5px;"> F2,S2 S3 </div> </div>	D2, S2, F2	MP2 Peak regional event upper greenschist to amphibolite facies. General recrystallisation overlaps D2.	
	D3, S3, F3	RMP3 Retrogression of MP2 assemblages.	
D4 <div style="display: inline-block; vertical-align: middle;"> { <div style="display: inline-block; vertical-align: middle; margin-left: 5px;"> S4a F4b </div> </div>	D4, S4, F4 D5, S5, F5	MP4 Biotite and chlorite porphyroblasts.	
DMG1		RMP4 Widespread retrogression.	Peak Thermal (Main Donegal Granite)
DMG2, SMG2, FMG2	D6, S6, F6		
?			
DMG3, SMG3	D7, S7, F7 D8, S8, F8 D9, S9, F9		

Table 2.2 Summary table of published chronology of metamorphism relative to deformation.

In a typical Barrovian type metamorphism a garnet-chlorite-muscovite-quartz assemblage could be produced by prograde metamorphism, but the pelites in the biotite and lower garnet zones may not carry biotite, because of the bulk composition being more appropriate to the production of magnesium-rich chlorites (Atherton 1964, 1965, 1968). If this is the case, then groundmass chlorites in pelites across Donegal may be earlier than those produced in the RMP_4 retrogression and there is evidence in central Donegal and Glencolumbkille of the retrogression of MP_2 biotite and garnet prior to the MP_4 growth of biotite porphyroblasts. If this is the case, then there may have been an early retrogression, RMP_3 (see Table 2.2) which affected regional chlorite-bearing assemblages instead of biotite-bearing rocks as at Glencolumbkille; this event is, however, very difficult to recognise given that it involves changes in chlorite only.

2.5.2 Contact metamorphism associated with the Thorr pluton

The thermal metamorphic event associated with the emplacement of the Thorr pluton has been described as a 'Static Recrystallisation' (Pitcher & Berger 1972; Oglethorpe 1987), that is low pressure, high temperature metamorphism, that produces a contact aureole up to 1.5 km wide. The aureole consists of mainly hornfelses that become increasingly mobilised on approaching the granite contact. It is notable that the thermal metamorphic event is increasingly reactive from the north of the pluton towards the south, coinciding with the lithological change from mainly quartzite in the north to predominantly pelite in the south. Detailed study of progressive mineralogical and textural changes is hampered on the regional scale by three factors: (a) the small fraction of pelite in the aureole (unreactive quartzite being in contact with the granite over much of the pluton); (b) the strike-parallel nature of most of the contacts and (c) the absence of any continuous tract of pelite going from the contact out into regionally metamorphosed country rock. Where pelite is juxtaposed next to granite, the 'static' nature of the aureole can clearly be seen, with andalusite developed in the outer aureole (e.g. at Gortahork) and sillimanite, cordierite and garnet developed in the inner aureole and rafts. Kyanite has also been reported from the southern part of the pluton by Edmunds (1969), but this complex area is within the aureole of the later Main Donegal and

Trawenagh Bay plutons, so it is difficult to assess whether kyanite is truly associated with Thorr and not these later plutons, which also resulted in thermal metamorphism.

Oglethorpe (1987), as part of his detailed study of the petrography and geochemistry of the granite near pelitic country rock, examined the thermal metamorphic assemblages in the pelite envelope and rafts. He subdivided the aureole into two, an inner (high grade) aureole and an outer (medium grade) aureole, and contrasted the mineralogy and texture of these zones of the contact aureole with the mineralogy and texture of the low grade pelites that had only been regionally metamorphosed. The following is a resumé of his study; the chronology of contact metamorphism to deformation events in the aureole will be discussed further in Chapter 3.

Inner aureole

The inner aureole can be studied at Curran's Point, Lettermacaward and in a continuous strip, parallel to the contact, from Crolly to Meencorwick (see Map 1). These hornfelses are generally sillimanite and cordierite bearing, but garnet is rare, which may be due, in part, to retrogression prior to granite emplacement. Staurolite is found at only a few localities and the poor occurrence of this mineral is believed to be due to overstepping of staurolite-forming reactions due to very rapid heating of the aureole hornfelses by the intruding granite (c.f. Naggar & Atherton 1970). At Curran's Point, in the north of the pluton, fibrolite occurs only in the immediate contact zone, within about ten metres from the contact. However, this situation is in strong contrast to the situation in the east and south of the pluton (Thorr and Lettermacaward - Maas), where there are kilometre-scale zones of thoroughly recrystallised rocks with an abundance of fibrolite. In the immediate contact zones in the east and south of the pluton, tourmaline is not uncommon; and in the south the growth of oligoclase has been noted adjacent to the contact in the so-called migmatitic rocks (Pitcher & Berger 1972).

In the inner aureole throughout the pluton, differential secondary recrystallisation has occurred in bands in which the segregation of quartz is strongly controlled by pre-existing planar structures. The segregation banding intensifies towards the contact, especially in the east and south of the pluton, where segregation is very well-developed, producing characteristic 'sweat-outs' of quartz, together with

concentrations of aluminosilicates, especially robust crystals of sillimanite, are found (Pitcher & Berger 1972).

A pronounced feature of the inner aureole is the mobilisation of the hornfelses, which is believed to be related in part to high pore fluid pressure during the thermal metamorphic event (see section 2.6.3) since it has been shown that prograde metamorphism can lead to high pore fluid pressure (mainly as a result of water liberated during dehydration reactions), which results in saturation of the rocks ($pH_2O =$ pload; Etheridge *et al.* 1983). If these events are then combined with suitably high geothermal gradients at depth, for example, adjacent to sites of granite ascent or emplacement, thermal expansion of water results, resulting in fracturing of the rocks (Norris & Henley 1976). In Thorr, Oglethorpe envisaged that high pore fluid pressure, exceeding the tensile strength of the rock, was generated in the pelitic layers by loss of water from muscovite and chlorite. This high pore fluid pressure in the plutonic environment could not be locally dissipated and thus resulted in a sudden loss of adhesion between grains in the pelitic layers. Such processes were relatively limited in the psammitic layers due to a lack of hydrous minerals prior to thermal metamorphism. As a consequence of this difference in behaviour, local overpressure occurred, resulting in fracturing of the more rigid quartzose layers and flow of the pelitic layers. This explains the common features of pieces of quartzite floating in pelitic horizons in the mobilised hornfelses and widespread quartzo-feldspathic veining seen throughout the Thorr area. Structure within the mobilised hornfelses of the inner aureole will be discussed further in Chapter 3.

Outer aureole

The outer aureole can be examined at Gortahork and Keeldrum (see Map 1) within the Falcarragh Pelite and Limestone Formations. These formations are separated from the granite by about 500m of quartzite. The constituent minerals of the hornfelses in this region are quartz, muscovite, biotite, andalusite, plagioclase, chlorite, ore and graphite. Decussate clusters of biotite, muscovite, iron ore and quartz are common and appear to represent thermally regenerated regional chlorite pseudomorphs of regional garnet (Oglethorpe 1987). The hornfelses of the outer aureole are finer-grained (approx. 0.1mm) than those in the inner aureole, except for andalusite and biotite porphyroblasts and occasional quartz stringers. There is a foliation in these hornfelses defined by muscovite and biotite, which according to

Rickard (1962, p.231) is related to the forceful emplacement of the Thorr pluton (see Chp. 3 for detailed discussion of this hypothesis).

All of the rocks in the outer aureole show metamorphic segregation banding, with muscovite and biotite being interbanded with granoblastic quartz. This is a feature which the pelitic and calcareous hornfelses of the outer aureole have in common with the pelitic and semipelitic hornfelses of the inner aureole.

Regionally metamorphosed, low grade pelites

These are easily observed at Ballyness pier and Falcarragh (see Map 1); they lie within the Falcarragh pelite formation. They have not been affected by the contact metamorphism associated with the emplacement of the Thorr pluton, showing evidence of only regional metamorphism at about 490Ma (Watson 1984). They are compositionally banded into quartz-rich and phyllosilicate (muscovite - chlorite) rich layers. There is a high angle between the compositional banding and the dominant cleavage which may indicate that although these rocks are moderately cleaved, the segregation may be partly due to an original sedimentary feature and not entirely due to pressure solution during cleavage formation (Oglethorpe 1987).

2.6 The Petrography and composition of the Thorr granite

The detailed petrography, composition or geochemical characteristics of the pluton, have been amply studied in previous research projects (Pitcher 1952; Whitten, 1956 & 1957 [see Pitcher & Berger 1972] Scott 1974 & Oglethorpe 1987) and much of the following text is a resumé of previous work.

The Thorr pluton is the oldest of the granites in NW Donegal, yielding an age of approximately 418Ma, which is believed to date intrusion (Rb-Sr whole rock; O'Connor *et al.* 1982). The pluton covers an area of approximately 800 km², most of which is submarine (See Fig. 2.8). The granite is mainly coarse-grained, frequently containing large alkali feldspar phenocrysts, which are often pink coloured. These and the mafic minerals form a moderately to locally strongly developed shape preferred orientation fabric, or magmatic state foliation (as defined in Ch.1) which is markedly bimodal in places (Whitten

1957 & Ch.4). This is often, but not always, parallel to the pluton margins (see Ch.4). These early magmatic state fabrics are variably overprinted by later solid state fabrics (*sensu* this thesis) which become progressively mylonitic towards the Main Donegal Granite Shear Zone (MDGSZ). In addition, there are narrow zones of fault breccia associated with approximately E-W trending, localised late fault zones (see Map 2).

Compositionally, the Thorr pluton is eccentrically zoned from margin to core (Pitcher & Berger 1972; Oglethorpe 1987). The central part of the pluton, the "Gola facies" of Whitten (1957b) (see Fig. 2.8), is rather felsic (see section 2.6.3 of this chapter) and locally very rich in alkali feldspar, although this often has a somewhat erratic distribution. The margins of the pluton are generally more mafic (granodioritic and dioritic), with relatively fewer alkali feldspar phenocrysts. However, the most basic rocks are found as xenoliths in the Main Donegal Granite and as a thin strip along the northern margin of the Main Donegal Granite (Pitcher & Berger 1972; Oglethorpe 1987). Throughout the pluton there is some evidence for coarser- and finer-grained sheets (Whitten 1957 & see Ch. 4), especially in the mainland coastal exposures east of the islands of Inishmeane and Inishirrer (Fig. 2.8 & Map 2). However, this study and previous work have been unable to delineate any clearly visible sharp break between the granitic central part of the pluton and the granodioritic and dioritic marginal facies in the field, although, as mentioned, local sheet-like internal contacts do occur (see Ch.3 for further details). In addition to these relatively impersistent internal contacts between granitic rocks of varying grain size, there are prolific dioritic enclaves found within the superficially homogeneous granite. One possible source of this very basic material is from appinitic bodies, of which there is a good example at Curran's Point on the north coast (Fig. 2.9 & Map 2). However, the obvious concentration of enclaves in the northern part of the mainland exposure of the pluton has led to speculation that there may have been several appinitic bodies along this coast in addition to that at Curran's Point, and that they became disaggregated upon granite intrusion (Pitcher & Berger 1972 & see Ch. 4 for further discussion). Another possible source of cognate xenoliths is from synplutonic dykes which have been disrupted; there is some evidence for this along the N and W coasts. However, it might be more realistic to view these two possibilities as part of the same process, since many appinitic bodies are closely associated with basic dyke swarms (c.f. Oglethorpe 1987

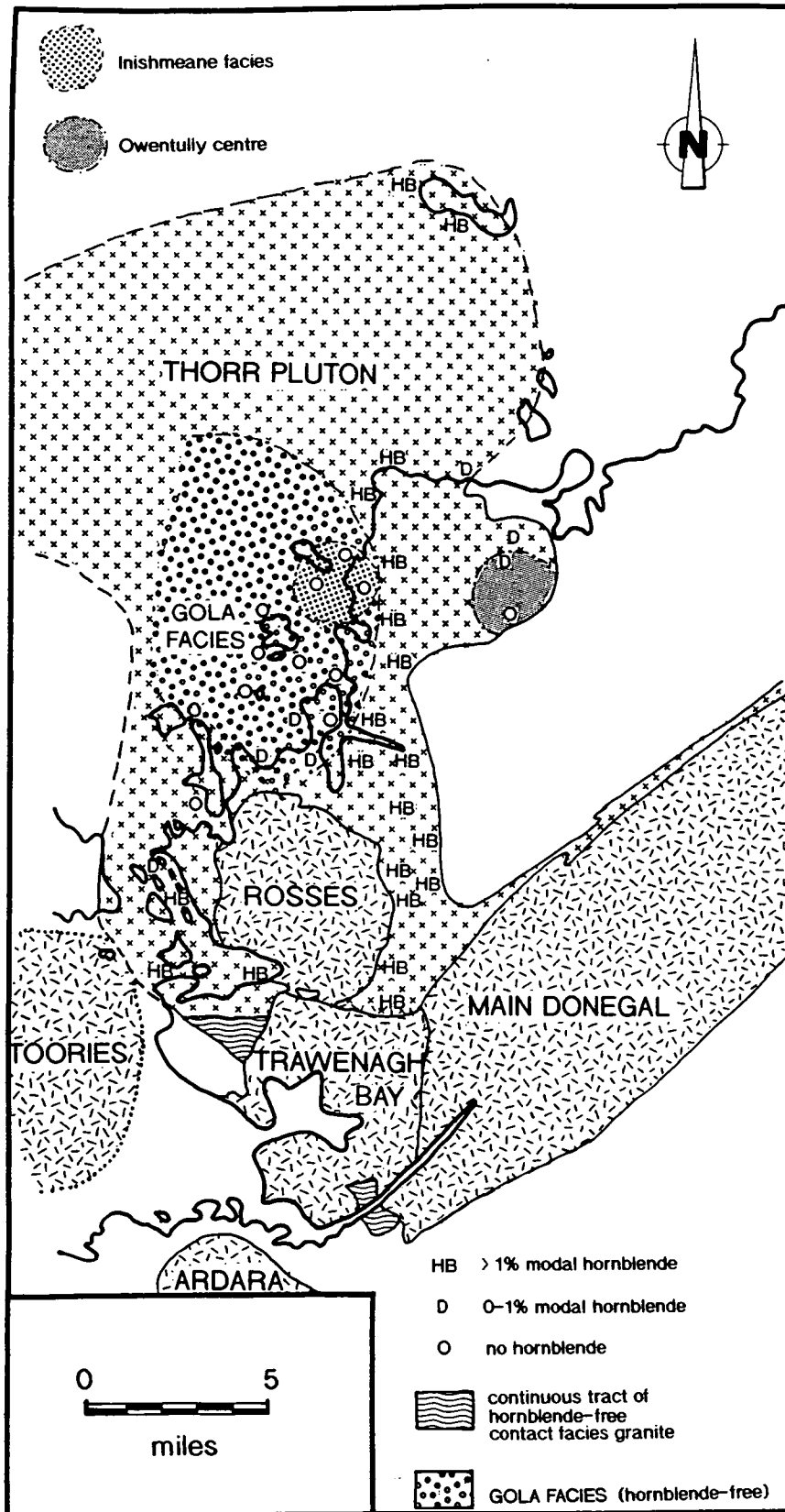


Figure 2.8 The Donegal granites, including the various facies of the Thorr granite (after Oglethorpe 1987).

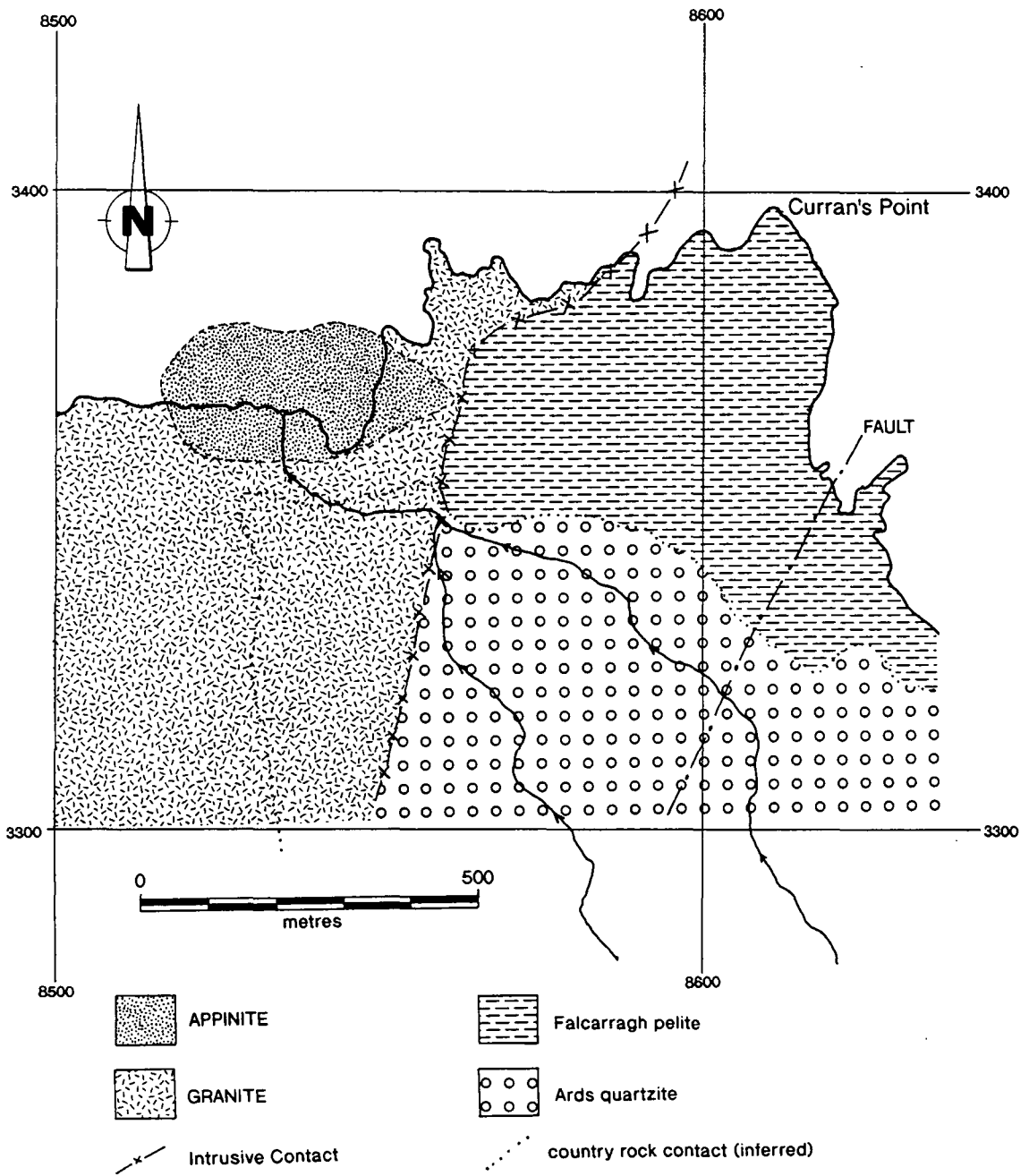
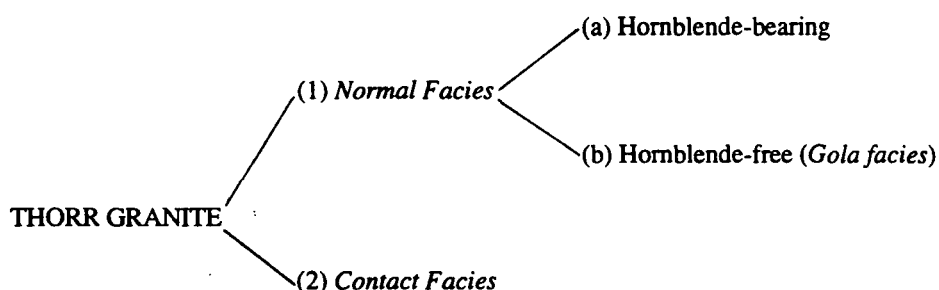


Figure 2.9 Map of the geology of the area around Curran's Point, showing the nature of the granite country rock contact in this area and the occurrence of an appinite body within the granite (after Oglethorpe 1987).

& see Ch. 5 on the Ratagain pluton). In addition to these cognate inclusions, there are also variably altered metasedimentary xenoliths, which range in size from a few mm to several tens of metres across and include the so-called rafts of the Thorr district (Maps 1 & 2).

The component facies of the Thorr pluton will be described in more detail in the following sections of this chapter. This involves subdivision of the pluton according to the modal occurrence of certain index minerals, for example hornblende, as well as details such as grain size. The result is that the pluton can be subdivided into the following constituent facies: *normal facies*, which is subdivided into *hornblende-bearing* and *hornblende-free (Gola facies)* components and *contact facies*.



The details of modal petrology and geochemistry are summarised after Pitcher & Berger (1972) and Oglethorpe (1987).

2.6.1 *The Normal facies*

This can be subdivided into hornblende-bearing and hornblende-free facies. The occurrence of the hornblende-free facies coincides with the area of the so-called 'Gola facies'. It is best to consider these facies separately, since they have slightly different characteristics.

2.6.1.a *Hornblende-bearing Normal facies*

This occupies the marginal parts of the compositionally zoned pluton (Fig. 2.8) and constitutes most of the present-day onshore outcrop. It ranges from quite mafic diorite, with colour index > 25, to granites *sensu stricto*, with colour indices around 10. Generally, however, colour index increases towards the pluton margins (Oglethorpe 1987). Constituent minerals include hornblende, plagioclase, biotite,

alkali feldspar and quartz; accessory phases include zircon, apatite, magnetite, ilmenite, allanite, sphene, pyrite, epidote and monazite. Oglethorpe (1987) determined the following order of crystallisation for the normal facies, hornblende-bearing Thorr granite, mainly on the basis of textural evidence:

EARLIEST	Zircon
	Apatite
	Magnetite/Ilmenite
	Allanite
	Sphene
	Hornblende
	Plagioclase
	Biotite
	Alkali feldspar
	Quartz
	Pyrite
LATEST	Epidote

This is an ordinary magmatic crystallisation sequence, typical of calc-alkaline plutons and it is common to the hornblende-bearing normal facies. Oglethorpe (1987) also reported that there is no petrographic or geochemical evidence to suggest that the normal facies granite was extensively contaminated by the country rocks, as envisaged by Pitcher and Berger (1972) (for further details of contamination process see section 2.6.3 of this chapter). The sequence of crystallisation is compatible with experimentally derived crystallisation sequences obtained from molten natural granodiorites, under water-deficient, vapour-present conditions *sensu* Robertson & Wyllie (1971a). The gradual changes in the proportion of phases (Whitten 1957b) and the decrease in M/FM^1 of hornblende and biotite and the anorthite content of plagioclase with the distance from the pluton margin suggests that crystal fractionation was dominant in producing the compositional zoning in this part of the pluton (see Oglethorpe 1987 for discussion of the detailed mechanism). The data are consistent with crystallisation of a granodiorite magma from the margins of the pluton inwards towards the core, in a manner similar to that proposed by Bateman & Chappell (1979) for the Tuolomne intrusives.

The majority of the constituent minerals in the granodiorite appear to have crystallised from a liquid with the exception of epidote and secondary sphene (Oglethorpe *op. cit.*). The cited author proposed that the textural features of these two minerals and their confinement to the southern part of the

¹ $M/FM = Mg^{2+} / (Fe^{2+} + Mg^{2+})$

pluton suggested that they were the product of either input of heat and volatiles from the later Trawenagh Bay and Rosses plutons, or of a higher subsolidus water fugacity in the south of the pluton due to water absorbed from the pelitic rafts and envelope rocks. However, it is impossible to determine precisely which mechanism was dominant, due to the coincidence of abundant pelitic rafts with the aureoles of the later plutons.

2.6.1.b *Hornblende-free Normal facies: The Gola facies*

This facies roughly occupies the centre of the pluton, but there is no obvious, well-defined contact with the surrounding hornblende-bearing normal facies. Oglethorpe (1987) proposed that this facies represents the end member of the continuous evolution of the Thorr magma during continued crystal fractionation. The rocks of this facies are all granites *sensu stricto*, with colour indices <10. There are some notable mineralogical differences from the hornblende-bearing normal facies:

- epidote and secondary sphene are absent and magnetite is the sole opaque phase. Primary sphene is considerably less abundant.
- biotite is browner and more strongly coloured.
- there are fewer zircon inclusions in biotite.
- plagioclase is more albitic (c. An₂₅) and more strongly optically zoned.

Both the hornblende-bearing and the hornblende-free facies have very similar grain size (1-2mm). However, in the north of Inishmeane Island and the adjacent mainland, a greyer, finer-grained (0.5mm) facies of hornblende-free granite, with prominent alkali feldspar megacrysts is seen (Inishmeane facies of Fig. 2.8). The modal composition of the finer-grained units is not anomalous (Whitten 1957b, p.274) and, generally all the normal facies components are typical of 'I' type granites (Chappell & White 1974). One possible explanation of the finer grain size in the Inishmeane facies is that they represent a late surge of felsic Thorr magma into the centre of the pluton (Oglethorpe 1987 & see Ch. 3 for further discussion).

2.6.2 *The Contact facies*

The term 'contact facies' is here used in the sense of Oglethorpe (1987) and describes a "granite of unusual mineralogy surrounding a particular lithology of border or raft." It has no genetic connotation. The contact facies varies, however, depending on whether it is in contact with the envelope rocks or country rock rafts. For the sake of clarity, the following subdivision will therefore be made in this thesis: *marginal facies* will be used to refer to granite with unusual mineralogy next to the border of the pluton, whereas *contact facies* will be used to refer to the zone of 'altered' granite around country rock rafts.

2.6.2.a *Marginal facies*

This can be seen in the pluton next to the envelope rocks, where it persists for approximately 10m after which there is a smooth gradation of characteristics into 'normal facies' granite. There is a general asymmetry to this marginal facies, with granodioritic rocks being found at the northern contacts (e.g. Bloody Foreland, Tory Island), where the granite is intruded dominantly into quartzite, and tonalitic and dioritic rocks occurring in the south of the pluton, where the granite is intruded into pelite and semipelite (Oglethorpe 1987). In hand specimen, the marginal facies varies from a pinkish, coarse-grained, alkali feldspar megacrystic, hornblende and sphene-bearing granodiorite, through to a white, equigranular, slightly finer-grained, biotite - muscovite tonalite. This facies often weathers to a rusty colour and occasionally contains small amounts of tourmaline and garnet, the latter being typical of crustally derived granites. In addition to muscovite, it also contains a greater modal proportion of quartz, and the plagioclase feldspars are more euhedral than in the 'normal facies'. There is no obvious change in colour index from the marginal facies to the normal facies. In general the marginal facies shows a tendency towards the characteristics typical of an 'S' type granite (Chappell & White 1974), especially in the southern part of the pluton (see section 2.6.3 of this chapter for details of the interaction between country rock and granite).

2.6.2.b Contact facies

This describes 'altered' granite around country rock rafts of pelite, semipelite and metadolerite. The actual mineralogy of the contact facies is different where it occurs adjacent to metadolerites compared to zones around pelites and semipelites, therefore, they will be considered separately. No contact facies 'aureoles' are found around the quartzitic rafts in the north of the pluton (for example at Bloody Foreland) suggesting that the composition of these rafts may be unsuitable for interaction with granite magma (Oglethorpe *op. cit.*). The contact facies in general is best developed around the large metasedimentary rafts in the areas of Ardveen, Meencorwick and Thorr, where it may be 5-10m wide. Scott (1974) and Oglethorpe (1987) charted in some detail the mineralogical changes that occur in the granite approaching a pelitic raft. The study highlighted the gradual nature of the change from granite with 'I' type characteristics, through a transitional facies ('T' type granite), through to granite with 'S' type characteristics (see Table 2.3 [after Oglethorpe 1987] and Fig.2.10 [after Scott 1974]).

Table 2.3 Typical sequence of mineralogical changes on approaching pelite contact

	NORMAL	<-----	TRANSITIONAL	----->	CONTACT
	'I'	'T ₁ '	'T ₂ '	'T ₃ '	'S'
Biotite colour	olive	olive	brown	foxy red-brown	foxy red-brown
hornblende	+	-	-	-	-
muscovite	-	-	-	-	+
sphene	+	+	±	-	-
epidote	+	+	±	-	-
allanite	+	+	±	-	-
magnetite	+	+	±	-	-
rutile	-	-	-	±	+
monazite	-	-	±	+	+

All facies contain plagioclase + quartz + alkali feldspar.

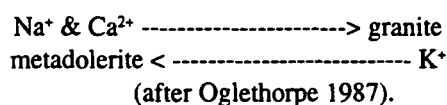
Like the marginal facies, there is no obvious change in colour index when compared with the normal facies. Petrologically, however, the contact facies is a white-coloured, biotite, muscovite tonalite. This is quite distinct from the marginal facies hornblende-bearing granodiorite. Once again, however, there is also an increase in modal quartz near the rafts and plagioclase is rather more euhedral than in the normal facies. The contact facies is slightly finer-grained (Oglethorpe 1987 & this study) possibly

suggesting that there has been some form of "chill" effect around the rafts of pelite and semipelite (see section 2.6.3 for discussion of interaction between pelite/semipelite and granite). Like the marginal facies, many of these characteristics are typical of 'S' type granites (Chappell & White 1974). Table 2.4 summarises the differences between the contact facies and the normal facies granite (after Oglethorpe 1987).

Table 2.4 Mineralogical differences between normal facies ('I' type) granite and pelite contact facies ('S' type) granite in the Meencorwick - Ardveen area.

	NORMAL FACIES	CONTACT FACIES
Hornblende, sphene, epidote, allanite	Present	Absent
Muscovite, monazite	Absent	Present (in inner part, closest to rafts)
Quartz content	Constant; average 15%	Erratic; average 15%
Microcline content	c. 15%	1-2% at main contact. Elsewhere, similar to normal facies
Plagioclase	Subhedral	Euhedral
Biotite colour	Olive	Foxy red-brown
Apatite	Relatively constant amounts; c. 0.3 vol.%	Variable amounts; 0.1-0.4 vol.%; cloudy quench cores
Pleochroic haloes	Small extent around zircon only	larger; found around zircon and apatite
Opakes	Euhedral magnetite, pyrite and ilmenite	Anhedral pyrite, rutile and ilmenite

Metadolerites produce a slightly different type of contact facies. It is a hornblende and sphene-rich, quartz-poor monzonite, which persists for only a metre or two around the metadolerites. In addition to the apparent change in mineralogy of the granite, the metadolerites have a thin, biotite-rich rim. Oglethorpe (1987) proposed that these changes could be produced by simply exchanging alkalis between the metadolerites and the granite as follows:



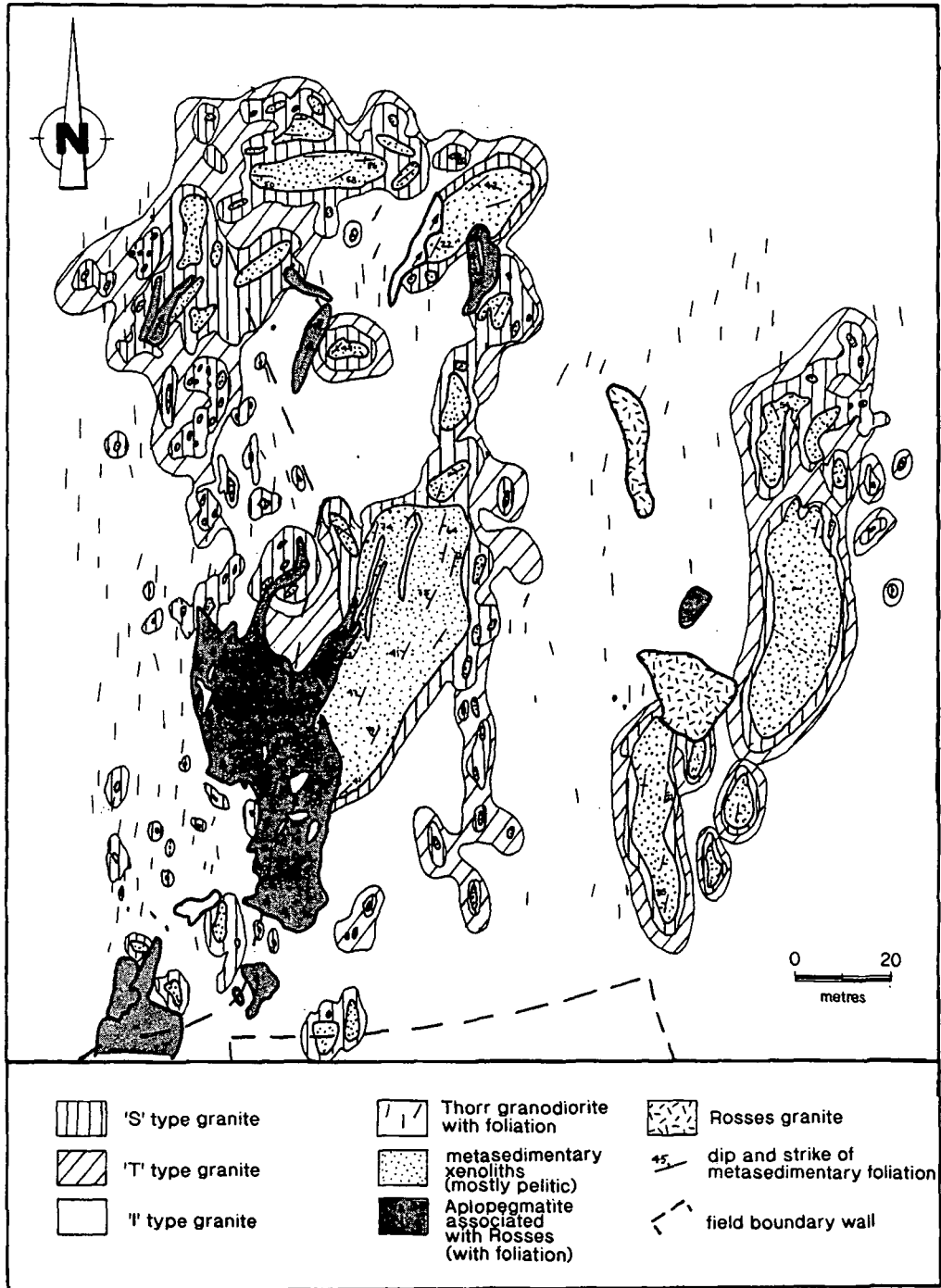


Figure 2.10 Map of the Ardeen area to illustrate the occurrence of I-, T- and S-type granites around pelitic country rock rafts (after Scott 1974).

2.6.3 Nature of the interaction between Thorr granite and its country rocks

Oglethorpe (1987) showed that whilst the interaction with pelitic metasediments had little effect on the mineralogy, chemistry and isotopic composition of the bulk of the Thorr pluton, the interaction between magma and pelitic country rocks resulted locally in very profound changes in both lithologies. This suggests that there was not wholesale contamination of the granite by partial melting and assimilation of country rocks. The only indication that partial melting of pelitic rafts and xenoliths might have occurred is provided by the presence of some hornfels showing segregation into mafic, biotite-rich layers and felsic, quartzo-feldspathic layers, but Oglethorpe (1987) did not believe this to be very significant and it may even have been an original metamorphic segregation banding which was accompanied by leaching of silica. He believed that evidence for the latter hypothesis came from the presence of abundant quartz veins in the pelitic hornfels.

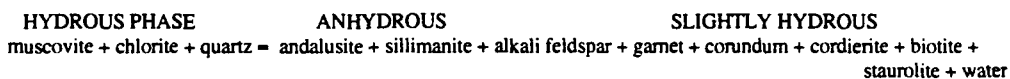
Textures in the inner aureole and the rafts are indicative of rapid heating of the pelites, inducing dehydration reactions, resulting in the sudden evolution of water. The above author also noted that p_{H_2O} in the aureole hornfels appears to have been buffered at around p_{TOTAL} . However, the data were determined by roughly calculating the compositions of mineral phases in equilibrium and using rough approximations of thermochemical data, fugacities of vapour components and activity relations of participating phases; in the absence of precise data for the conditions listed above the values obtained are, at best, a rough guide. The mineralogy and mineral chemistry of the pelitic rafts, however, are indicative of p_{H_2O} being less than p_{TOTAL} at the peak of thermal metamorphism, suggesting extraction of water into the undersaturated granite. Oglethorpe (*op. cit.*) postulated that the pelite rafts were never subjected to any significant degree of partial melting because water had been effectively 'sucked' out of the pelite rafts. Furthermore, as he believed that this part of the pluton was near the roof zone, since the rafts are thought to be roof pendants, heat loss would have been rapid. He suggested that within the envelope pelites, p_{H_2O} probably never fell below p_{TOTAL} and heat could then be effectively dissipated by convective processes driven by high fluid pressure and temperature gradient (c.f. Etheridge *et al.* 1983), thereby preventing the rocks from ever reaching their solidus temperature. Oglethorpe (1987) also noted widespread evidence of deformation suggesting that both the rafts and the envelope rocks became highly

plastic and mobilised as a result of interaction with the hot granitic magma, this he attributed to the effects of turbulent convection. Earlier deformation structures (c.f. section 2.4) are often still recognisable, in addition to rather disharmonic folds formed during the mobilisation event (for further discussion see Ch.3).

Major and trace element analyses of granites and raft pelites (Oglethorpe 1987) indicate that alkalis were also mobile during interaction. The process of ionic diffusion is thought to have led to the apparent movement of Ba^{2+} , Sr^{2+} and K^+ into the raft pelites and Rb^+ into the magma. CH_4 and silica were also seemingly carried in solution from the rafts to the magma. Analysis of aureole pelites, however, suggests that alkalis did not diffuse into these, possibly because the large amounts of water which moved across the main contact 'swamped' out any diffusive movement of ions and appears to have leached soluble cations out of the marginal magma.

According to Oglethorpe (*op. cit.*) changes in the characteristics of the pelitic country rocks due to interaction with the Thorr granite magma can be summarised as follows:

- (1) rapid growth of new thermal metamorphic minerals (see Chp. 3 for details)
- (2) loss of water (as indicated by the net reaction)



- (3) mobilisation of the pelite and leaching of silica
- (4) progressive loss of graphite (Atkin 1977).

Likewise, proximity of the granite to pelitic material resulted locally in major changes to the magma, particularly in mineralogy, which can be related to movement of water and ions with resultant changes in water vapour pressure and oxygen fugacity. As mentioned previously, the granite in the south of the pluton contains secondary sphene and non-magmatic epidote, which may be partly attributed to higher subsolidus pH_2O in this region due to water being absorbed from the pelitic rafts and envelope. In addition, the reversal in the order of crystallisation of plagioclase and biotite (illustrated in Figure 2.11) suggests, according to Oglethorpe (1987), that pH_2O was considerably higher in the marginal facies compared to the contact facies around the rafts (Robertson & Wyllie 1971a, p.272) The presence of more

numerous fluid inclusions in marginal facies apatites also supports this hypotheses. Magmatic ilmenite is present only in the southern raft-rich part of the pluton, being absent from rocks of similar SiO₂ content in the north of the pluton. Oglethorpe (*op. cit.*) suggested that this may have been a response to lower oxygen fugacity in the south of the pluton, possibly because of the widespread reducing effect of the pelitic rafts.

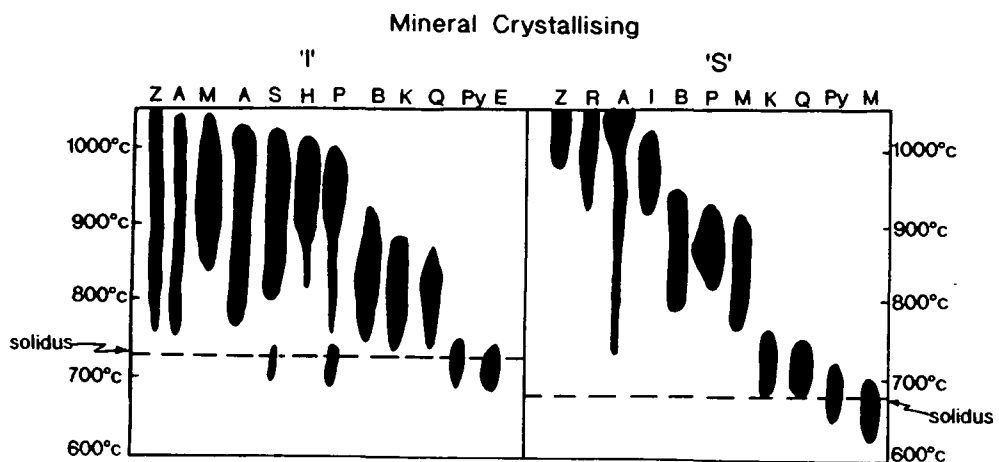
Figure 2.11

(a) In order of commencement of crystallisation

'I'	'S'
Zircon	Zircon
Apatite	Rutile
Magnetite/Ilmenite	Apatite
Allanite	Ilmenite
Sphene	Biotite
Hornblende	Plagioclase
Plagioclase	Monazite
Biotite	Alkali feldspar
Alkali feldspar	Quartz
Quartz	Pyrite
Pyrite	Muscovite
Epidote	

(a) Mineralisation of a normal facies ('I') and contact facies ('S') granite in the Thorr pluton (after Oglethorpe 1987)

(b) Order of crystallisation and amount of crystallising mineral as a function of temperature



(b) Estimate of the amount of crystallisation of individual minerals as a function of temperature in 'I' and 'S' type Thorr magmas (after Oglethorpe 1987; drawn with reference to Robertson & Wyllie 1971a). Thickness of black shaded area is representative of the amount of individual minerals crystallising at a given temperature.

Heat gain in the pelitic rafts appears to have been balanced by heat loss in the magma, as suggested by the presence of quench apatite in the marginal and contact facies (Oglethorpe 1987). If one reasonably assumes that all processes of heat loss would be at a maximum early in the emplacement history, when the temperature difference between the magma and the country rocks was greatest, it is possible to envisage rapid cooling of the pluton, especially at its margins from the granodiorite liquidus of 1050°C to a more stable temperature of around 800°C (Oglethorpe *op. cit.*). However, the granite at the contact cannot have been completely quenched since it is still fairly coarse-grained, so the overall effect of the absorbed water was that it acted like a flux, lowering the solidus temperature and extending the temperature interval over which crystallisation would have taken place. This factor is extremely important to bear in mind when considering the fabric development in the pluton (see Ch.3).

Mineralogical changes in the characteristics of the granite due to interaction with pelitic country rocks can be summarised as follows:

- (1) disappearance of hornblende and appearance of muscovite
- (2) change in biotite colour and chemistry
- (3) increase in the modal proportion of SiO₂
- (4) increase in the plagioclase/alkali feldspar ratio at the main contact
- (5) crystallisation of a different set of opaque and accessory minerals
- (6) growth of quench apatite containing fluid inclusions.

Fluxes of H₂O, CH₄, SiO₂ and alkalis across contacts also seems to have had a secondary effect on the granite chemistry by changing the mineral stabilities in the melt, leading in turn to enrichment of the contact facies granites in P⁵⁺ and depletion in light rare earth elements (LREEs) and Thorium (Oglethorpe 1987).

One of the main conclusions of Oglethorpe's work was that the changes in the mineralogy of the early crystallising accessory minerals and the presence of quench apatites at country rock contacts, in addition to variations in the mineral chemistry, suggest that the Thorr magma was virtually crystal-free when it was initially emplaced. This is a very important point to note, since it has profound implications

on the rheological response of the magma if it was subjected to any form of stress during emplacement, particularly tectonic stresses (see Ch.4 for further discussion).

2.7 Events post-dating the emplacement of the Thorr pluton

Granitic and microgranitic sheets, dykes and plugs outcrop extensively in the southern part of the Thorr pluton. These are associated with the earliest phases of intrusion of the Rosses pluton (Pitcher & Berger 1972; Oglethorpe 1987). They increase in number and thickness towards the contact between the Thorr and Rosses plutons, and are cross-cut by the main outcrop of the latter body. Chemical analysis confirm the comagmatic nature of these minor intrusions with the Rosses granite (Oglethorpe 1987). In addition, aplite and pegmatite sheets and dykes, possibly also related to the Rosses granite, occur throughout the Thorr pluton.

Emplacement of Thorr was followed closely in time, according to the most recent geochronological studies (e.g. O'Connor *et al.* 1987) by the intrusion of the Rosses Central Complex. It is seen in contact with the older granodiorite in the south - central part of the Thorr pluton. It seems to produce only a localised aureole within the host granodiorite (Pitcher 1953a). At Lough Anure and Lough Nanillan (Maps 1&2), however, limestone xenoliths within the Thorr granite have wide alteration rims of skarn minerals as a result of the proximity of the more water-rich Rosses magma (Oglethorpe 1987).

The intrusion of Rosses was immediately succeeded by the intrusion of the Main Donegal Granite and the Trawenagh Bay granite. The former has been very well-studied (e.g. Hutton 1982a and references therein) and is known to be intimately associated with sinistral shearing along the Main Donegal Granite Shear Zone (MDGSZ). In contrast, the Trawenagh Bay pluton is poorly understood, but appears to cross-cut all the afore mentioned plutons. O'Connor *et al.* (1987) suggested that the whole of the Donegal Batholith was intruded in a period of only 20Ma.

Movement on the MDGSZ continued after the emplacement of the Main Donegal Granite (Hutton 1982a). Both the early and late movements on the MDGSZ are recorded in both the MDG and

the southern part of the Thorr pluton. Deformation associated with ductile movements along this shear zone produced sinistral s-c fabrics in the SE part of Thorr, which overprint the deformation fabric associated with the emplacement of the Thorr pluton itself (see Chs.3&4).

On cooling, the Thorr pluton developed a joint pattern dominated by a set oriented NNW-SSE. The granite has also been fractured in places by cataclastic shear zones, possibly associated with the MDGSZ or later brittle movements in the same orientation, i.e. ENE-WSW (see full description of cataclasites and gouges in Ch.3).

During or soon after emplacement the Thorr pluton appears to have been rapidly uplifted and eroded, since the K-Ar dates on biotites and hornblende are approximately 380Ma (Brown *et al.* 1968), almost within the error of the Rb-Sr isochron. Unfortunately, in the Thorr area there are no remnants of this period in the geological record. In fact, the only rocks of post-Caledonian age, prior to Pleistocene and recent deposits, are NW-trending Tertiary Olivine Dolerite dykes which are particularly well exposed along the north coast, near Bloody Foreland (see Map 2).

A layer of boulder clay was deposited during the Pleistocene, particularly in the north of the pluton. In addition, Recent peat deposits are also developed in this area. Together they have a cumulative thickness in the range of tens of metres in the north of the pluton, resulting in poor inland exposure along the north coast.

Chapter 3
**COUNTRY ROCK DEFORMATION AND CONTACT RELATIONS IN THE
REGION OF THE THORR PLUTON**

Introduction

The previous chapter outlined the published schemes of deformation, metamorphism and granite intrusion events in NW Donegal. In this chapter, structural and petrographical data will be presented from the country rocks surrounding the Thorr pluton. The nature of the contacts between the Dalradian metasediments and the granite will also be described. The data from the country rocks and the contacts will then be interpreted in order to determine those structural and metamorphic features that are related to the emplacement of the granite, and those which pre- or post-date intrusion.

The present author has used the deformation chronology defined by Rickard (1962, 1963) and Hutton (1979, 1982a), to constrain the relative timing of the emplacement of the Thorr granite. Within the existing chronology, the Thorr pluton is believed to have been emplaced after D_3 but before D_6 ; therefore, it pre-dates the main phases of deformation associated with the emplacement of the Main Donegal Granite (see Table 2.1). However, instead of using the sometimes confusing 'D' numbering system, deformation structures will be described in terms of 'early', 'main' and 'late' phase structures, with cross-references to published regional events where appropriate. The structural, metamorphic and petrological data has been subdivided into various sub-areas (see Fig. 3.1), according to the geographical location, country rock lithology and style of intrusion and deformation. Within each sub-area, the most obvious set of folds are designated as the 'main' phase, with all other structures defined as 'early' or 'late' relative to the 'main' phase folds. The sub-areas studied are listed below (for geographical locations see Fig. 3.1):

- (1) Binanea Strand - Ards Point, Tory Island (see inset to Map 2), Bloody Foreland Mountain and Tievelehid;
- (2) Curran's Point and Inishbofin Island;

- (3) Aranmore Island;
- (4) Maghery;
- (5) Lettermacaward;
- (6) the country rock rafts at Lough Agher and Ardveen;
- (7) Crollly - Thorr - Meencorwick.

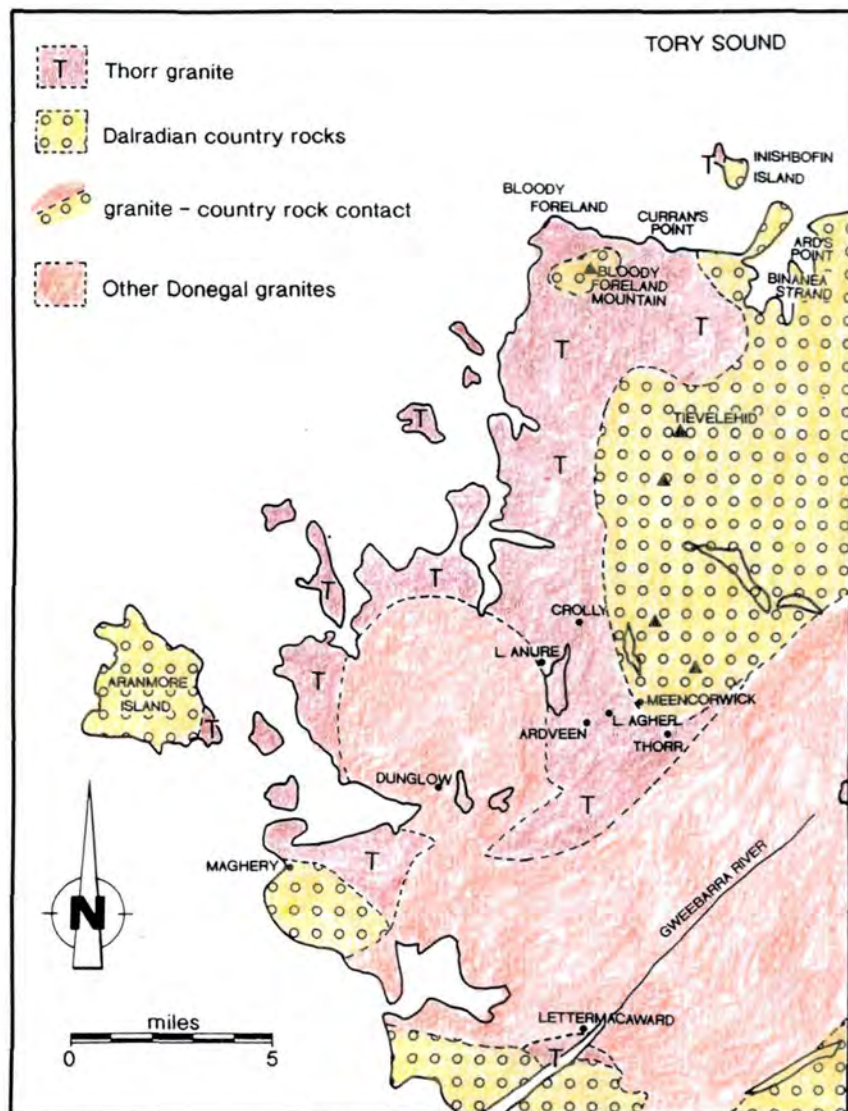


Figure 3.1 General map of the geology of NW Donegal showing most of the study areas referred to in this chapter (for the location of Tory Island see Map 2).

3.1 *Field observations and analysis of country rock structure and contact relationships*

3.1.1 *Binanea Strand - Ards Point and associated areas*

3.1.1.a *Binanea Strand - Ards Point*

Whitten (1957) has described the nature of the quartzitic country rocks and the contact with the granite in this area. He referred to the country rocks as migmatised Ards quartzite with lit-par-lit granite veins. The present study has found no evidence for migmatisation of the quartzite in this area. The geometry of the bedding-parallel granite veins (see below), however, could broadly be described as lit-par-lit, although not in the sense used to describe migmatites, since there is no evidence of melting in the country rocks.

The country rocks are well-bedded on a cm- to m-scale, with obvious cross-laminations and cross-bedded structures indicating that the strata are right way up. The beds mostly dip shallowly towards the SE-SSE (see Fig 3.2a). A well-developed stretching lineation is seen on the bedding surfaces, shallowly plunging down the dip of the bedding surfaces, (see Fig. 3.2b). It is defined by elongate quartz and feldspar grains, and mullion structures are often aligned parallel with the lineation. There are no observable mesoscopic fold structures in this area.

The texture of the quartzite has been examined in thin sections. It is best described as semi-annealed, with a strong preferred orientation of elongate quartz grains and rounded feldspar grains (see Plate 3.1).

Several sets of granite veins are seen in this area (see Fig 3.2c);

Set 1: are roughly parallel to the bedding in the quartzite, ranging in width from a few centimetres to a maximum of 0.5 metres (c.f. Fig. 3.2ci);

Set 2: are shallowly inclined towards the NW-NNW and range in width from a few centimetres to tens of centimetres (c.f. Fig. 3.2cii);

Set 3: are mostly steeply SE-dipping veins, that range in width from a few centimetres to tens of centimetres (c.f. Fig. 3.2ciii).

BINANEA STRAND - ARDS POINT

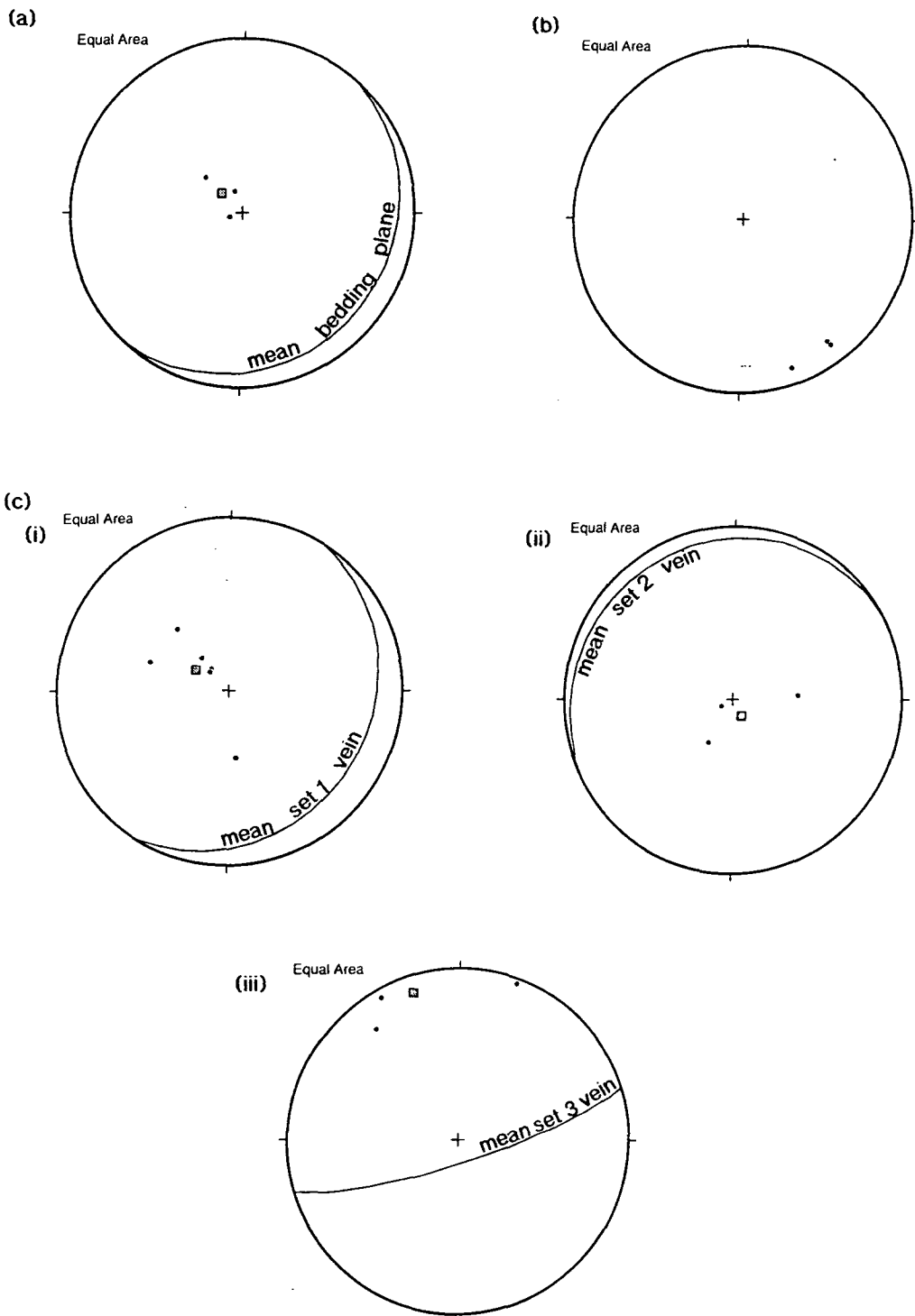


Figure 3.2 Equal area stereonets of structural data from Binanea Strand. (a) poles to bedding planes; (b) mineral lineations; (c) granite veins, (i) set 1, (ii) set 2, (iii) set 3.



Plate 3.1 Photomicrograph of Ards Quartzite collected from close to the granite contact at Binanea strand. The texture is semi-annealed, and quartz and feldspar show a well-developed preferred orientation. (Field of view 18mm).



Plate 3.2 Photomicrograph of Ards Quartzite collected from close to the northern contact with the Thor Granite on Bloody Foreland Mountain. Feldspar is relatively abundant in the quartzite here, a fact that has been attributed to metasomatism close to the roof zone of the pluton. The texture of the quartzite is semi-annealed, and constituent minerals exhibit a well-developed preferred orientation. (Field of view 18mm).

The texture of all three vein sets varies from aplitic, through granitic to pegmatitic; the thickest veins, which are usually those that are parallel with the bedding in the quartzites, tend to be the most pegmatitic. The opening direction for all sets of veins has been defined by offsets of markers in the quartzites and the direction of growth of crystals within the granite veins. In all cases, there is no evidence of any lateral offset of markers across the granite veins, and the direction of grain growth is always perpendicular to the vein walls. Both of these facts are consistent with orthogonal opening of the vein sets, suggesting that they formed as tensile (Mode I) fractures. The significance of this mode of intrusion will be discussed fully in section 3.3.2b.

3.1.1.b *Tory Island*

Whitten (1957a) described the country rocks and contact relationships on this island. The current study reveals that on the island, the granite is in contact with well-bedded, right way up Ards Quartzite, based on the widespread preservation of cross laminations. Beds dip shallowly to moderately towards the east (see Fig.3.3a). As at Binanea Strand, there is a well-developed stretching lineation defined by alignments of quartz, feldspar, and sometimes muscovite. The lineation plunges shallowly to the SSE (see Fig. 3.3b), and often millimetre structures are aligned parallel to it. There are no observable mesoscale deformation structures, such as folds, exposed on the island.

A limited number of veins were observed around the island, mainly in the area around Port Doon on the east of the island (see Map 2, NC 877455). All veins that intrude the country rock are tensile fracture infills, as determined from lack of lateral offset of markers across the veins. On Tory Island, however, unlike the first set of veins at Binanea Strand, the granite veins are not exactly parallel with the bedding, striking clockwise at about 70° of bedding, and having mostly shallow NNW dips (see Fig. 3.3a &c). To the west of Arderill (see Map 2), the contact is exposed in a vertical sea cliff, where it appears to be sharp and trends NW-SE; there is no inland exposure. In the southeast of the island, in the flat tidal pavement near Torranaman (see Map 2), the sharp contact is rather irregular and transgressive with respect to the bedding in the quartzite country rocks. South of the main contact, randomly orientated xenoliths of quartzite are enveloped in coarse-grained granite; similar xenoliths are also seen at

TORY ISLAND

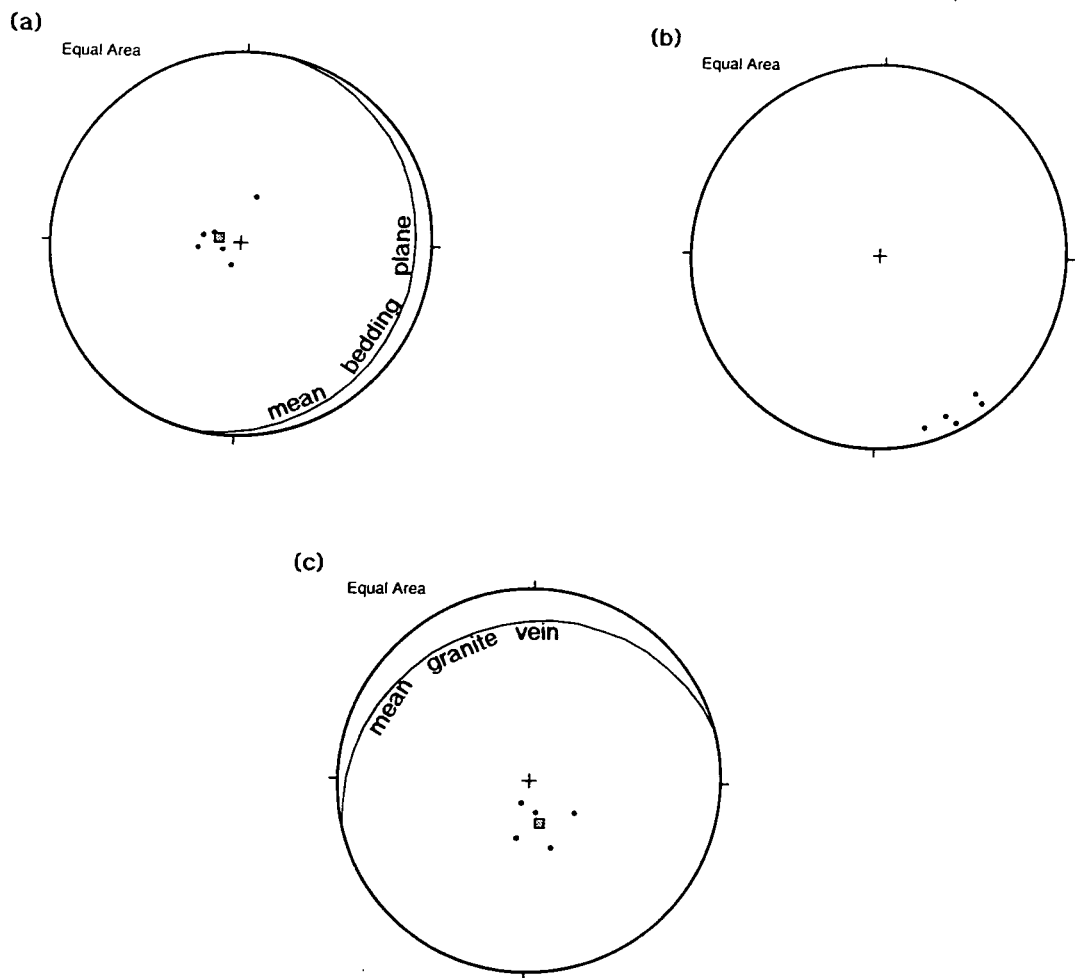


Figure 3.3 Equal area stereonets of structural data from Tory Island. (a) poles to bedding planes; (b) mineral lineations; (c) poles to granite veins.

Rinardallif Point, near Bloody Foreland on the mainland (see Map 2). Immediately to the east of the contact at Torranaman, the quartzite is cut by several large, shallowly inclined granite sheets and a network of interlocking microgranite dykes and sills.

3.1.1.c *Bloody Foreland Mountain*

At Bloody Foreland Mountain (see Map 2) a quartzite 'cap' sits on top of the hill overlying granite. Whitten (1957) described the petrography of this contact in detail, referring to it as 'steep'. This is not quite consistent with the observations of the present study, which suggests that the granite contact is sharp and roughly parallel to the bedding in the quartzite, dipping moderately to steeply in variable directions from NE through to SE; ESE-SE is the average. This relationship is observed everywhere along the exposed contact, except in the extreme west of the section (see Map 2), where the granite contact cuts across the quartzite bedding in an E-W direction. The contact here is moderately inclined towards the south, with the granite dipping underneath the quartzite. It is roughly parallel to gently S-dipping granite veins observed up to 50 metres into the quartzite. Whitten (1957) reported that many granite-quartzite contacts in the area are transitional, but this was not noted during the present study, since all observed granite contacts are sharp. The only noticeable change in the quartzite close to the granite is its colour, which becomes pinker. Close inspection of hand specimens reveals that the pink colour is due to the increased occurrence of feldspar in the quartzites close to the contact. This has been confirmed in thin section (see Plate 3.2). This is perhaps indicative of limited metasomatism in this area, perhaps in response to a high concentration of volatiles within the roof zone of the pluton. The texture of the quartzite from Bloody Foreland is rather annealed in thin section, but with a very strong shape preferred orientation (Plate 3.2). The annealed texture is typical of a rock that has been significantly hornfelsed in a thermal aureole.

3.1.1.d *Tievelehid*

The granite margin with the country rocks is not well-exposed here due to a thick covering of boulder clay and peat. The contact is exposed approximately 200m due south from Lough Lagha (see

Map 2) in a stream section. The contact is sharp and dips moderately towards the NW, which is approximately parallel to the strike and dip of the well-bedded, right way up (from cross-laminations) quartzite country rocks. The quartzite is baked in this area, but it does not contain as much new feldspar as the quartzite near the contact at Bloody Foreland and, therefore, does not have such a distinctive pink colour. Whitten (1957) described a 'migmatitic' (lit-par-lit) contact zone almost due north of Tievelehid summit, but this author could find no evidence of such contact relations. No granite veins were seen in this area, a fact that may be due, at least in part, to the very poor exposure.

3.1.2 *Curran's Point and Inishbofin Island*

3.1.2.a *Curran's Point*

Here, the granite is seen along a coastal section, in contact with pelites that are believed to belong to the Falcarragh pelite formation (see Map 2 & Fig. 2.2; Whitten 1957; Pitcher & Berger 1972; Oglethorpe 1987). A compositional interbanding of pelitic and semipelitic horizons is clearly identifiable in the field. This banding is clearly folded (Fig. 3.4a) around centimetre- to metre-scale main phase folds that coaxially refold locally developed earlier folds that have a strongly transposed crenulation cleavage parallel to their axial planes. A stretching lineation, defined mainly by phyllosilicate minerals, is also folded around the hinges of the main set of folds, producing a wide scatter of the lineation data on the stereonet (Fig. 3.4b). In the field, the main set of folds is developed on a metre-scale, with parasitic folds developed on a centimetre-scale. Interlimb angles are variable, but the average is close to tight. These folds have a moderate- to strongly-developed, millimetre-spaced crenulation cleavage subparallel to their axial planes which generally dip steeply NW-NNW (Fig. 3.4c). This main set of folds mostly plunge at moderate angles to the NE, with a few plunging to the SSW (Fig. 3.4d). Where they are seen to be curvilinear in the field, they have a whaleback geometry, with hinges passing through the strike of their axial planes. In the field, these folds appear to consistently verge up to an antiform in the WNW-NW and, as the mean plunge is towards the NE, this observation is consistent with the fact that the axial planes to these folds generally strike clockwise to the foliation. To the west of the section, near the contact with the granite, the country rocks have a very platy appearance, which is indicative of high

CURRAN'S POINT

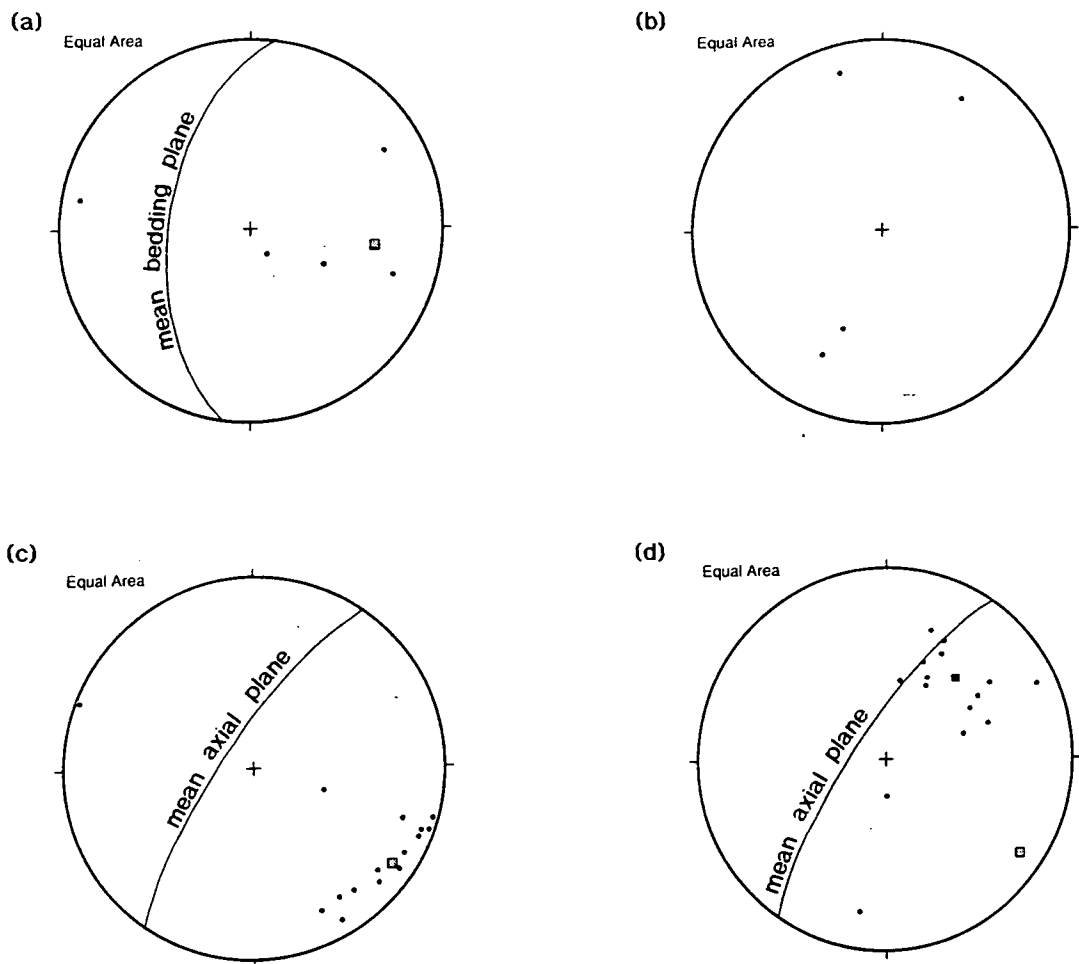


Figure 3.4 Equal are stereonets of structural data from Curran's Point. (a) Poles to bedding; (b) stretching lineation; (c) poles to fold axial planes; (d) fold axes.



Plate 3.3 Platy rocks adjacent to the contact with the Thorr Granite at Curran's Point. These are believed to have developed in response to high strains prior to granite emplacement.



Plate 3.4 Randomly orientated country rock xenoliths within a granite vein at Curran's Point. The lack of orientation of xenoliths is interpreted as being consistent with a rather passive mode of emplacement of these veins.

strains (Plate 3.3). Whilst the high strain is believed, in part, to pre-date the main phase of folds, since the platy fabric is folded, the platy zones also coincide with the long limbs of the main phase folds, suggesting that strain was perhaps also quite high during and subsequent to the formation of these folds. Granite veins in the field can be seen to cross-cut the main phase folds and to be intruded roughly parallel to the axial planar crenulation cleavage, although they sometimes dip in the opposite direction. The contacts between granite veins and the pelitic country rocks are always sharp, although, it is common to find variably digested xenoliths of country rock within the granite veins. The xenoliths of country rock tend to be randomly orientated within the veins (see Plate 3.4), suggesting that granite emplacement was rather passive at this level, because, if it had been associated with active flattening, one would expect to see folding and/or alignment of the country rock xenoliths. Contact metamorphism resulted in the growth of biotite and fibrolitic sillimanite in the pelitic and semipelitic country rocks at Curran's Point; these minerals are randomly orientated and overprint the crenulation cleavage associated with the main fold phase.

3.1.2.b *Inishbofin Island*

Here the granite is seen in contact with compositionally banded pelites and semipelites. Rickard (1962) called these rocks the Inishbofin banded semipelite/psammite group, but the present author cannot find any criteria with which to distinguish them from the banded pelites and semipelites exposed at Curran's Point and it is, therefore, suggested that they are equivalent. In the field, the foliation/compositional banding generally dips moderately towards the NW along a NE-SW strike (see Fig. 3.5a). The compositional banding is folded by one obvious set of main phase folds on the island. There is a well-developed, roughly axial planar, spaced crenulation cleavage associated with this set of folds. The intensity and spacing of the cleavage is variable, but the average spacing is on a scale of about 0.5mm. In several observed localities west of Inishbofin School (see Map 2), the crenulation cleavage intensifies within the hinge zone of some of the folds and small thrust faults with a few tens of centimetres of displacement are developed. The small thrust faults are usually associated with quartz veins about 50cm wide. Stereonet (b) in Figure 3.5 summarises the field data relating to the axial planes, crenulation

INISHBOFIN ISLAND

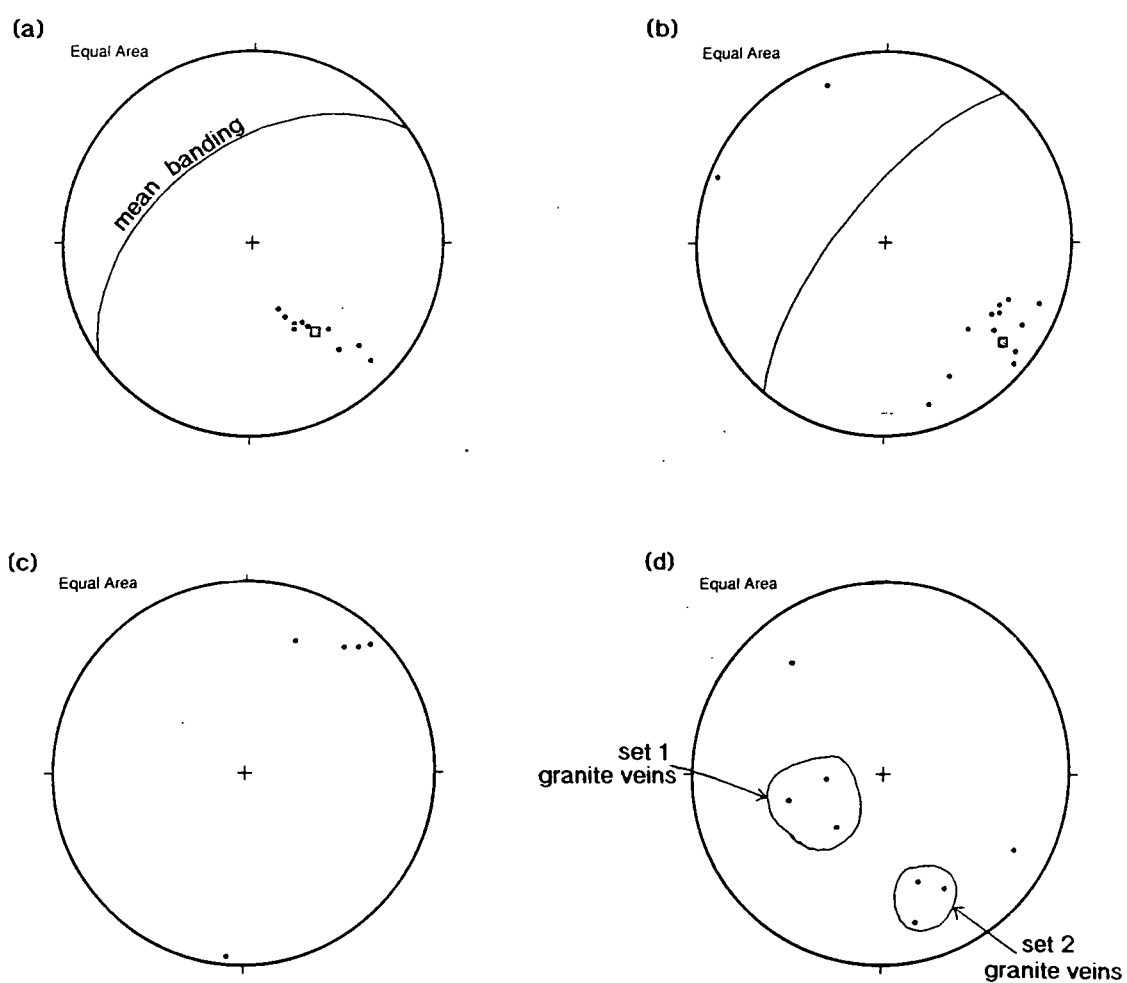


Figure 3.5 Equal area stereonets of structural data from Inishbofin Island. (a) Poles to banding; (b) poles to fold axial planes and crenulation cleavage; (c) fold hinges; (d) poles to granite veins with subsets labelled.

cleavage and thrust faults associated with the main phase folds on Inishbofin Island. The hinges of these folds, where seen, generally plunge shallowly towards the NE (Fig. 3.5c); they are not obviously curvilinear in the field. The majority of the main phase folds verge up to an antiform in the SE, therefore synforming to the NW (see Map 2). At the NE end of the island, the granite margin appears to roughly coincide with the axial trace of a putative mesoscale synform, based on the presence of a large zone of neutral vergence (\approx synform hinge) seen at Toberglassan Bay (see Map 2). This zone of neutral vergence is cut out by the granite southwestwards and is not preserved in the south of the island.

The granite contact is exposed in coastal sections in the S and NE of the island (see Map 2). Where exposed it is always sharp. At a few localities near to the granite contact in the south of the island (e.g. NC 888358), sinistral shear bands were seen to have developed subparallel to the crenulation cleavage.

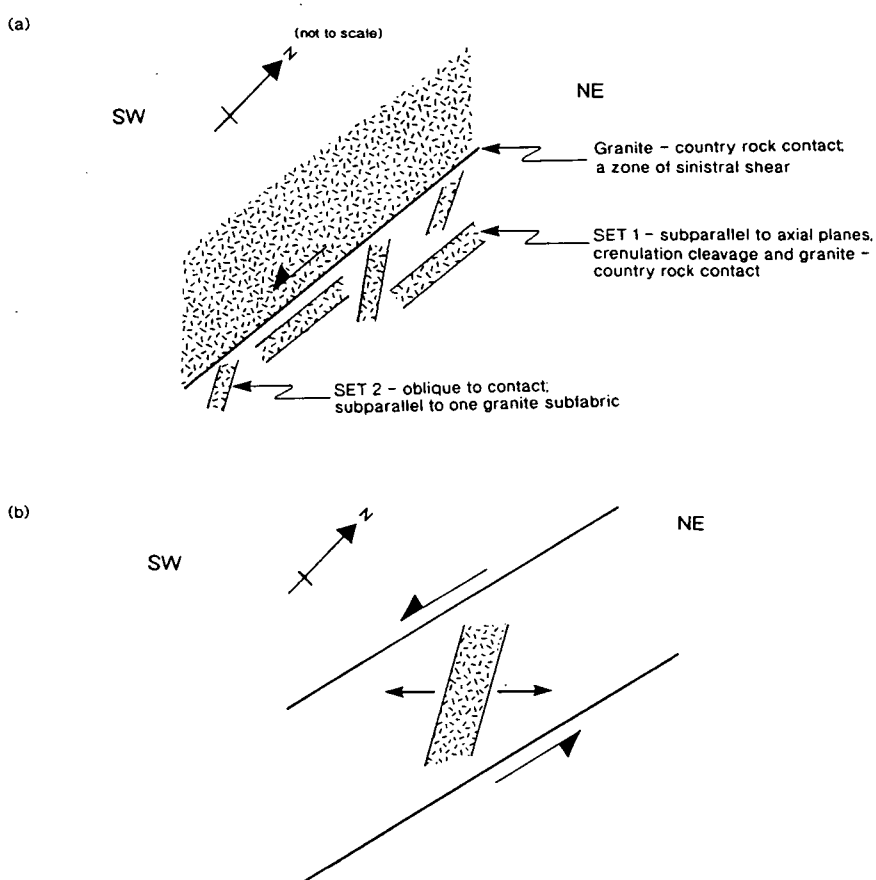
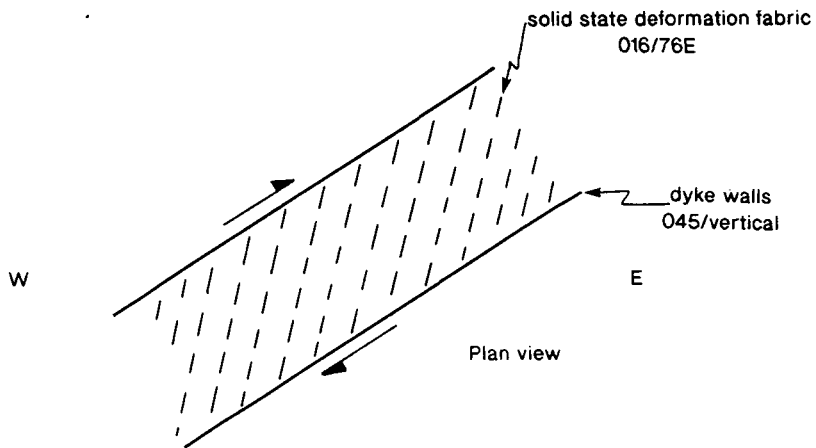


Figure 3.6 Granite veins on Inishbofin; (a) sketch summarising the orientation of granite veins relative to the main contact; (b) theoretical opening direction and orientation of granite veins under conditions of sinistral shear parallel to the main granite - country rock contact on the island.

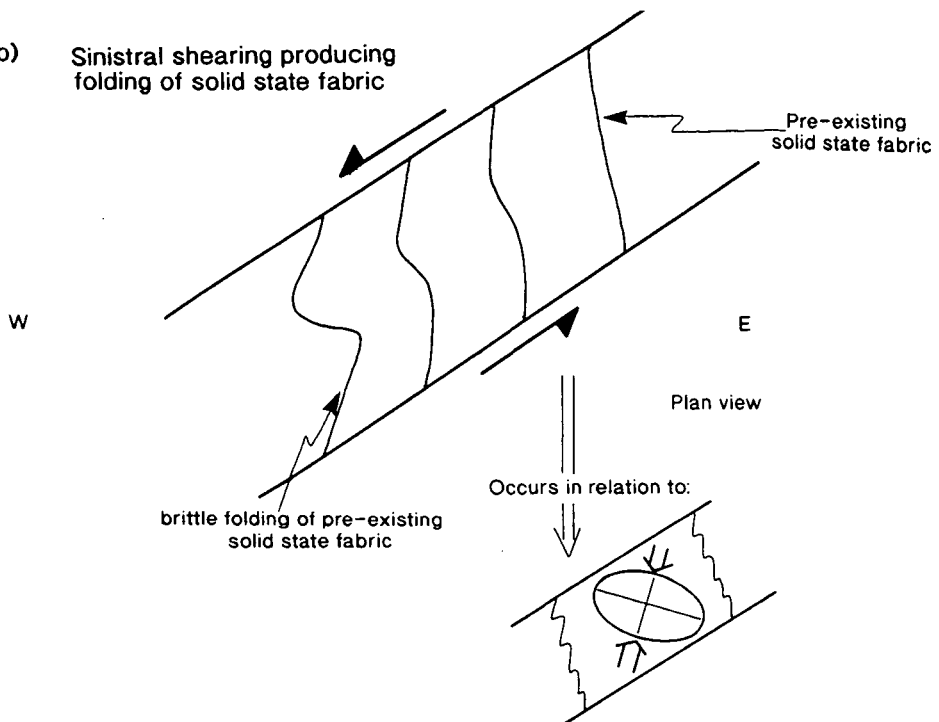
Two sets of granite veins are observed on the island (Fig. 3.5d), but there are no cross-cutting relationships that could be used to determine their age relative to one another. All contacts between granite and country rocks are sharp. One set of granite veins lies approximately subparallel to the axial planes of the main fold phase and the crenulation cleavage, that is they strike NE-SW and dip NW. The second set lie anticlockwise of the first (as illustrated in Fig. 3.6a), trending approximately N-S to NNW-SSE; this orientation is subparallel to one of the subfabrics seen in the granite on this island (see Ch. 4). The opening vectors of the second set of granite veins are consistent with a component of sinistral shear parallel to the granite contact, as illustrated in figure 3.6b; it is also consistent with shear sense indicators within the granite (see Ch. 4).

Additional evidence of sinistral shear at, or close to, the granite contact on the island is seen at Scolthageeragh, just to the SE of Erroonagh Bay (see Map 2), where low temperature mylonites and cataclastically deformed granite is intruded by metre-scale, relatively undeformed lamprophyre dykes. These are also believed to be Caledonian in age since they are petrographically similar to the synplutonic dykes seen within the Thorr granite along the N and W coasts. The granite at Scolthageeragh carries mm- to cm-scale brittle fractures with mm- to cm-scale sinistral offsets. Within this fault zone, the granite is rich in mafic minerals, mainly hornblende retrogressed to biotite. In addition to sinistral movements, there is some evidence of extensional movements along fractures within this fault zone. Extensional fractures are also seen to offset granite pods in dioritic veins within the fault zone. Microgranite dykes at Gobrinatroirk also carry evidence of sinistral shearing. Steeply plunging 'brittle' S-folds here deform an earlier dextral solid state deformation fabric within the microgranite veins, as illustrated in figure 3.7. All the above observations suggest that this particular region of the granite contact formed an area of long-lived deformation involving mostly sinistral displacement, possibly with some phases of additional dextral shear throughout the Caledonian orogeny and early Caledonian plutonism.

(a) Dextral solid state fabrics in microgranite dyke



(b) Sinistral shearing producing folding of solid state fabric



(not to scale)

Figure 3.7 Foliations in microgranite veins on Inishbofin Island. (a) Formation of dextral solid state deformation fabric during dextral shearing and its orientation relative to intrusion walls; (b) folding of solid state deformation fabric due to later sinistral shearing.

3.1.3 Aranmore Island

Exposure of country rocks is mainly limited to the north of the island (see Map 2) whilst granite contacts are poorly exposed in the south of the island. Hence, the information outlined below is based on only a few outcrops. The country rocks on the island are mainly quartzites, correlated with the Ards quartzite by Pitcher & Berger (1972). In the south of the island, however, semipelitic rocks are exposed near the contact with the granite. It was impossible, due to lack of exposure, to ascertain whether these are interbedded semipelitic units within the Ards Quartzite, or whether they represent other more pelitic successions, such as the Falcarragh pelites. In addition, a few metadolerite units are exposed on the island in association with the quartzites.

The metasediments, particularly the semipelites, exhibit a compositional banding, striking between ESE-WNW and E-W. At one locality (see Map 2, NC 682151), semipelites were folded by one obvious set of close to tight, centimetre- to metre-scale folds. These folds plunge moderately towards the towards the W and verge up to an antiform in the N. No way up evidence was observed. The folds are associated with a moderately developed, axial planar, mm-spaced crenulation cleavage, striking ESE-WNW and dipping moderately to steeply towards the SSW. Anastomosing, cm-wide granite veins are found throughout this outcrop. These granite veins have variable strikes, ranging from NW to NE, dipping moderately from NE to W. The semipelites at the east of this outcrop are in contact with coarse-grained porphyritic Thorr granite; this contact trends approximately NNW-SSE, but it was not possible to measure the dip. The granite here carries a well-developed magmatic state deformation fabric trending NW-SE, defined mainly by feldspar phenocrysts.

In the south of the island another contact between the granite and the country rocks is exposed (see Map 2, NC 679149). The granite margin here strikes roughly ESE and dips steeply towards the NNE. The strike of this contact is sub-parallel with, but dips steeply in the opposite direction to the crenulation cleavage seen elsewhere on the island. However, this crenulation is not particularly well-developed at this locality, perhaps due to baking and annealing of the country rocks in the thermal aureole of the Thorr Granite. There is some evidence of shearing of the country rocks at this locality, suggested by deflection of the foliation. The sense of shear is, however, ambiguous, with one cm-wide shear having

sinistral deflections and another having dextral deflections. The apparent sinistral shear strikes approximately E-W and is sub-vertical. The sinistral and dextral shears are never seen cross-cutting one another, so it is not possible to say anything about their relative timing. The dextral shear is also steeply dipping, but trends SE-NW. These features are only exposed at one locality, so it is difficult, therefore, to assess the significance of these structures with regard to the emplacement of the Thorr granite.

Country rock xenoliths within the granite on Aranmore Island always trend E-W. The nature of the fabric within the granite will be discussed in Chapter 4, but it is relevant to note here that there are two subfabrics present.

3.1.4 *Maghery*

The structure of the country rocks in the Maghery area was studied most recently by Meneilly (1982); all other studies preceding this are summarised in Pitcher & Berger (1972). A short section from the N side of Maghery Bay towards the granite at Termon Headland (see Map 2) was studied by the present author. The country rock lithologies are interbanded semipelites and calcareous pelite horizons, believed to be part of the Falcarragh Pelite succession that pass into a contact facies of the Thorr granite. The foliation in the semipelites, dips steeply on average towards the SW and SSW (Fig. 3.8a). A stretching lineation defined mainly by phyllosilicate minerals is developed within the semipelites and is widely scattered on a stereonet (Fig. 3.8b) due to folding about an axis that plunges shallowly towards the ESE. Early boudinage structures are developed in green-coloured calcareous pelite horizons, lying almost orthogonal to the stretching lineation where they are exposed together. The data are consistent with synchronous development of the stretching lineation and boudinage. The boudin necks appear to act as focii for the development of many main phase disharmonic folds.

Two sets of folds are observed in the Maghery area within the semipelitic horizons: the main set of folds coaxially refold early folds, producing Type III (hook) interference patterns (Ramsay 1967).

The early folds

The early folds consistently verge towards an antiform in the N-NE, although they are only preserved in a few places, mainly in the long limbs of the main folds. Where preserved, they occur on a

MAGHERY

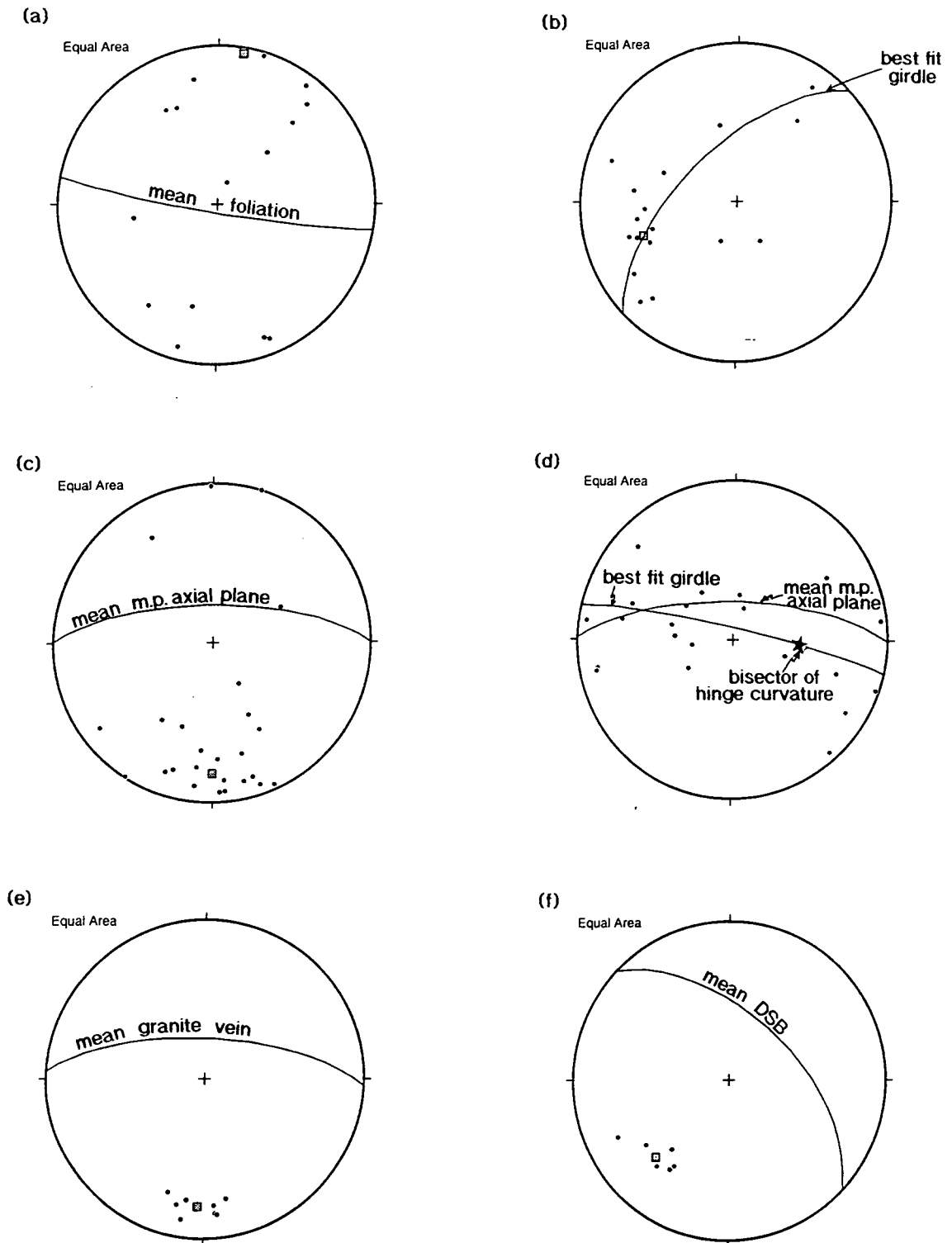


Figure 3.8 Equal area stereonets of structural data from Maghery. (a) Poles to foliation; (b) mineral lineations; (c) poles to main phase fold axial planes; (d) main phase fold axes; (e) poles to granite veins; (f) poles to dextral shear bands.

centimetre- to metre-scale, and are tight to isoclinal structures (see Plate 3.5). Their axes lie roughly parallel to the stretching lineation and they fold an earlier bedding-parallel fabric.

The main folds

The axial planes of the main set of folds generally strike E-W and dip moderately to steeply towards the N (see Fig. 3.8c). There is some spread of data which is believed to relate to the curvilinear nature of the axial planes which can be directly observed in the field. Axial planar cleavage is only occasionally observed in association with the main folds, possibly due to the effects of annealing in the thermal aureole of the granite (c.f. Meneilly 1982). When plotted on a stereonet, the fold axes to the main phase folds are spread along a sub-vertical, ESE-WNW girdle (Fig.3.8d). This reflects the curvilinear nature of the folds, which is clearly seen in individual exposures. In most cases, they are curvilinear through the strike of their axial planes (i.e through the horizontal), varying from gently ESE-plunging, to steeply W-plunging (see Fig. 3.9). Locally, however, the plunge steepens sharply to pass through the dip of their axial plane (i.e. through the 90° pitch in that surface) to plunge NE (see Fig. 3.9).

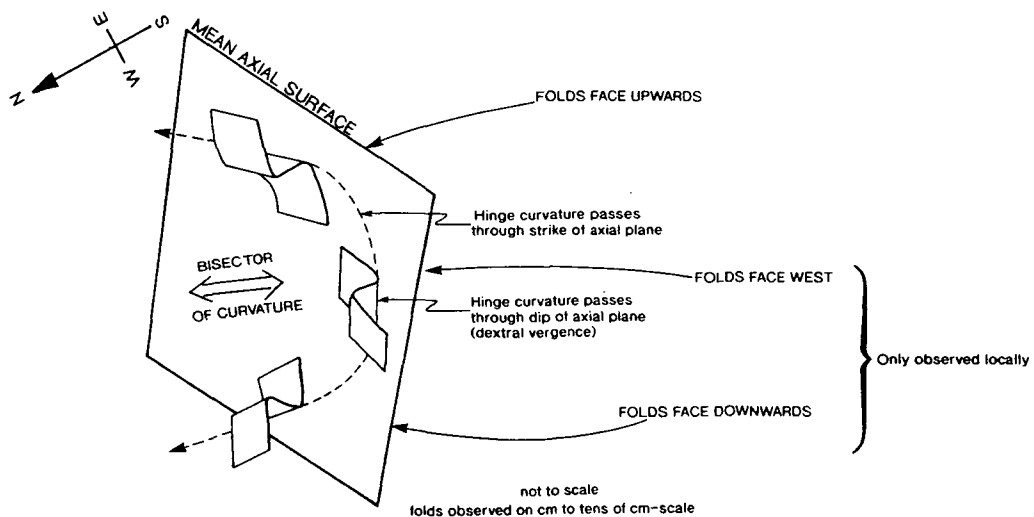


Figure 3.9 Sketch to illustrate the geometry and vergence patterns of the main phase folds at Maghery.



Plate 3.5 Example of a tight to isoclinal, decimetre-scale early fold in semi-pelites and calc-pelites at Maghery.



Plate 3.6 Examples of over-tightened main phase folds in semi-pelitic lithologies close to the contact with the Thorr Granite at Maghery.

When the main phase folds pass through the dip of their axial planes they exhibit dextral vergence as illustrated in figure 3.9. The bisector of the locus of fold hinges seen at Maghery is marked on stereonet 3.8d; it has a moderate plunge towards the east and should represent the stretching direction during the formation of this sheath fold phase (Holdsworth & Roberts 1984). Thus the mineral lineations that are folded by the main phase folds are clearly earlier and are partially reorientated towards the later finite X axis of bulk strain.

Strain appears to increase rapidly towards the granite contact in the north of the section, as does the occurrence of granite veins. The increase in intensity of strain coincides with an increase in the tightness of the main phase folds on a metre- to tens of metres-scale (see Plate 3.6). In the 10 - 20 metres immediately south of the granite margin, veins can be observed with a spacing of only a few centimetres. The veins are between a few mm and tens of cm wide. Contacts between granite veins and the country rocks are always sharp, although there is some very localised evidence of melting in some semipelites. The granite veins are irregular and do not carry a particularly strong fabric, either in the field, or in thin section, suggesting that they have not been emplaced in association with particularly intense strains. There is no evidence in this area to suggest that the high strain zone in the contact region country rocks was in any way related to the emplacement of the Thorr granite. It is, therefore, the opinion of this author that the high strain zone and the main fold phase with which it is associated, had formed before granite emplacement occurred. This interpretation concurs with that of Meneilly (1982), who stated that, in his opinion, the Thorr granite at Maghery was intruded passively, not forcefully as proposed by Pitcher & Berger (1972). The former author also stated that the steep dip of the foliation near the contact was not the result of oversteepening in relation to granite intrusion, but that it was a result of the pluton contact being exposed within the vertical hinge zone of the Maghery anticline. It is conceivable that the zone of tight to isoclinal main phase folds nearest to the granite lie in the hinge zone of an anticline, as no consistent sense of vergence could be determined within the high strain zone (Fig. 3.10), however, the present author cannot conclusively state that this is equivalent to the Maghery Anticline as described by Meneilly (1982).

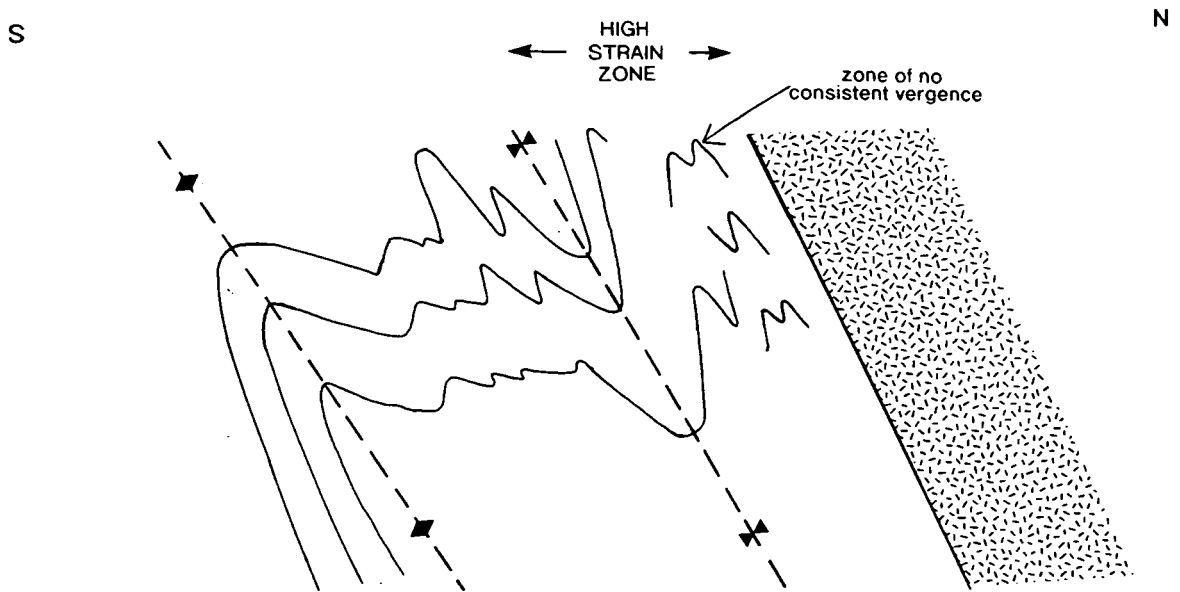


Figure 3.10 Sketch to illustrate steepening of the foliation and lack of consistent fold vergence within the high strain zone close to the granite contact at Maghery.

Granite veins in this area have a dominant E-W strike, and dip steeply to the N (Fig. 3.8e), unlike the foliation, which dips predominantly southwards. On average, the granite veins are subparallel with the main fold axial surfaces, but within the section, granite veins are locally seen to cross-cut the main fold axial planes.

The fabrics within the granitic rocks will be discussed in detail in Chapter 4, but generally they carry a weak to moderate magmatic state fabric, defined by a preferred orientation of feldspar and mafic minerals. Occasionally, however, the granite veins carry a moderately-developed solid state fabric, which can sometimes appear to 'undulate', as if it had been buckled. Open, irregular buckling of thinner granite veins is not uncommon within the contact zone, the veins being apparently compressed in a SSW-NNE orientation. All these features, together with the earlier country rock structures, appear to be overprinted by dextral shearing, with mm- to cm-spaced dextral shear bands occurring both in the country rocks and within the granite veins. The dextral shear bands have steep NE dips and NW-SE strikes (see Fig. 3.8f)

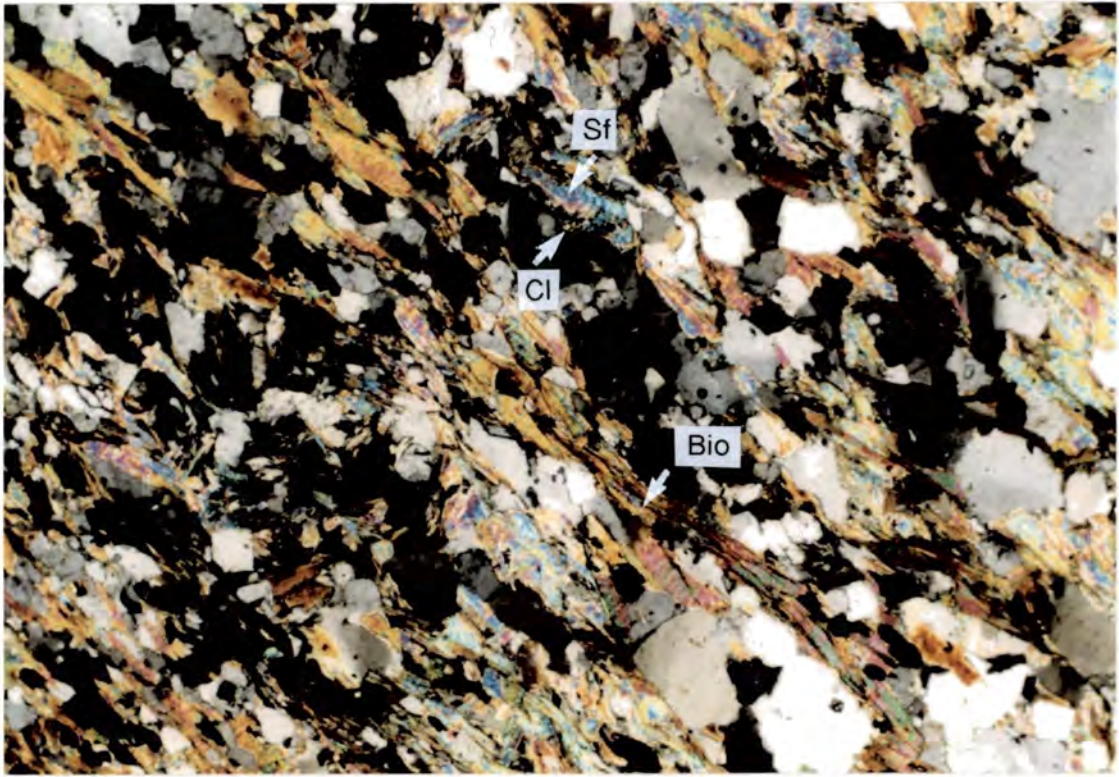


Plate 3.7 Photomicrograph of hornfelsed country rocks collected from close to the granite contact at Maghery.

Field observations suggest that the hornfelses in this area have experienced some dextral shearing *after* the thermal metamorphism related to granite emplacement. The kinematics of deformation cannot be confirmed in this section. However, the development of chlorite after sillimanite and especially biotite at points of open 'kinking' in relation to this late deformation event is consistent with deformation at relatively low temperatures and pressures.

Cl = Chlorite

Bio = Biotite

Sf = Sillimanite

(Field of view 4.4mm).

and lie clockwise of the foliation and of the granite veins (Fig. 3.8a, e & f), which is consistent with dextral shearing. In thin section, the dextral shear bands are associated with the growth of chlorite after fibrolitic sillimanite and biotite, indicating that they are rather low-grade features (see Plate 3.7). The brittle-ductile nature of the shear bands, as observed in the field, is confirmed in thin section, where they can often be seen to fracture quartz and feldspar grains, but appear to also exhibit ductile deformation features, such as undulose extinction and subgraining of quartz grains (see Plate 3.7). This dextral shearing event is also manifest in the granite to the E and SE of Maghera Bay and to the west of Lough Illion (see Map 2). In these areas it results in the development of a very weak solid state deformation fabric, mainly manifest as a flattening of quartz grains (see Ch.4).

3.1.5 *Lettermacaward*

Pitcher & Berger (1972) described this area of the Thorr granite as a thoroughly migmatized and mobilized zone with numerous country rock xenoliths (mainly pelite). The main contact zone and the country rock rafts within this two-mica facies of the Thorr granite were examined in a short reconnaissance visit during this study. The pelitic country rocks carry a well-developed, mm-spaced crenulation cleavage. Biotite and sometimes sillimanite, are seen to replace muscovite along the cleavage planes. The general structure in the country rocks is reminiscent of the structure in the pelitic and semipelitic lithologies elsewhere around the pluton (see below). There is one main phase of close to tight fold structures in the area, developed on a scale of centimetres to tens of centimetres. The crenulation cleavage mentioned above is axial planar to these folds and it dips moderately towards the NNE and strikes ESE-WNW. The folds observed during this brief study generally plunged gently towards the east, and generally have 'Z' shapes looking down the plunge of the fold axes, suggesting that the folds verge up to an inclined antiform in the S and synform to the N. The observed folds refold an earlier mineral lineation, defined by muscovite, around their hinges.

Later structures within the pelites in this area include infrequent millimetre- to centimetre-wide sinistral shear bands, with only a few millimetres offset, that are filled with granite. These trend roughly E-W and dip very steeply towards the N. In addition, occasional cross-cutting granite veins, centimetres

to tens of centimetres wide are intruded into the pelitic country rocks and rafts. These granite veins do not carry a well-developed fabric. They strike ESE-WNW, subparallel to the axial planes of the folds in the area, however, they dip steeply in the opposite direction (i.e. towards the SSW).

3.1.6 *The rafts at Lough Agher and Ardveen*

3.1.6.a *Lough Agher*

Scott (1974), as part of a geochemical study, mapped one large raft (approx. 250m x 900m), and several smaller rafts in this area. Lithologically, they are an interbanded mixture of pelites and semipelites, with minor amounts of calc-pelite. The total outcrop of rafts in this area is no more than 600m x 900m, so the structural data were collected in a relatively small sample area. The rafts are partly mobilised, although not as intensely as the 'migmatites' that will be described in section 3.1.8 of this chapter. In terms of their mineralogy, they are best described as sillimanite-cordierite-garnet schists, the metamorphic assemblage clearly reflecting intense contact metamorphism. Sillimanite often occurs as the variety fibrolite pseudomorphing andalusite and, on rare occasions, andalusite cores are preserved, with fibrolite replacing the outer parts of the euhedral andalusite.

The rafts carry a well-developed compositional banding. This generally strikes NNE-SSW, the angle dip is variable, but is generally towards the SE. The data are distributed along a girdle, the pole of which roughly corresponds to the mean fold axis of the main folds observed in the area (Fig. 3.11a & f). The strike of the mean foliation in the Lough Agher area is slightly clockwise of the foliation seen in the migmatites to the east (see Map 1 & compare Figs. 3.11a and 3.14a). This author interprets this to be a result of slight rotation of the rafts away from the main granite - country rock contact.

There are no well-developed stretching lineations developed in the Lough Agher area, but occasionally a weak mineral lineation, defined mainly by phyllosilicates, is seen to plunge shallowly towards the SE-SSE (Fig. 3.11b).

Two sets of folds are developed in this area; an early set and a main set, the latter being most common.

LOUGH AGHER

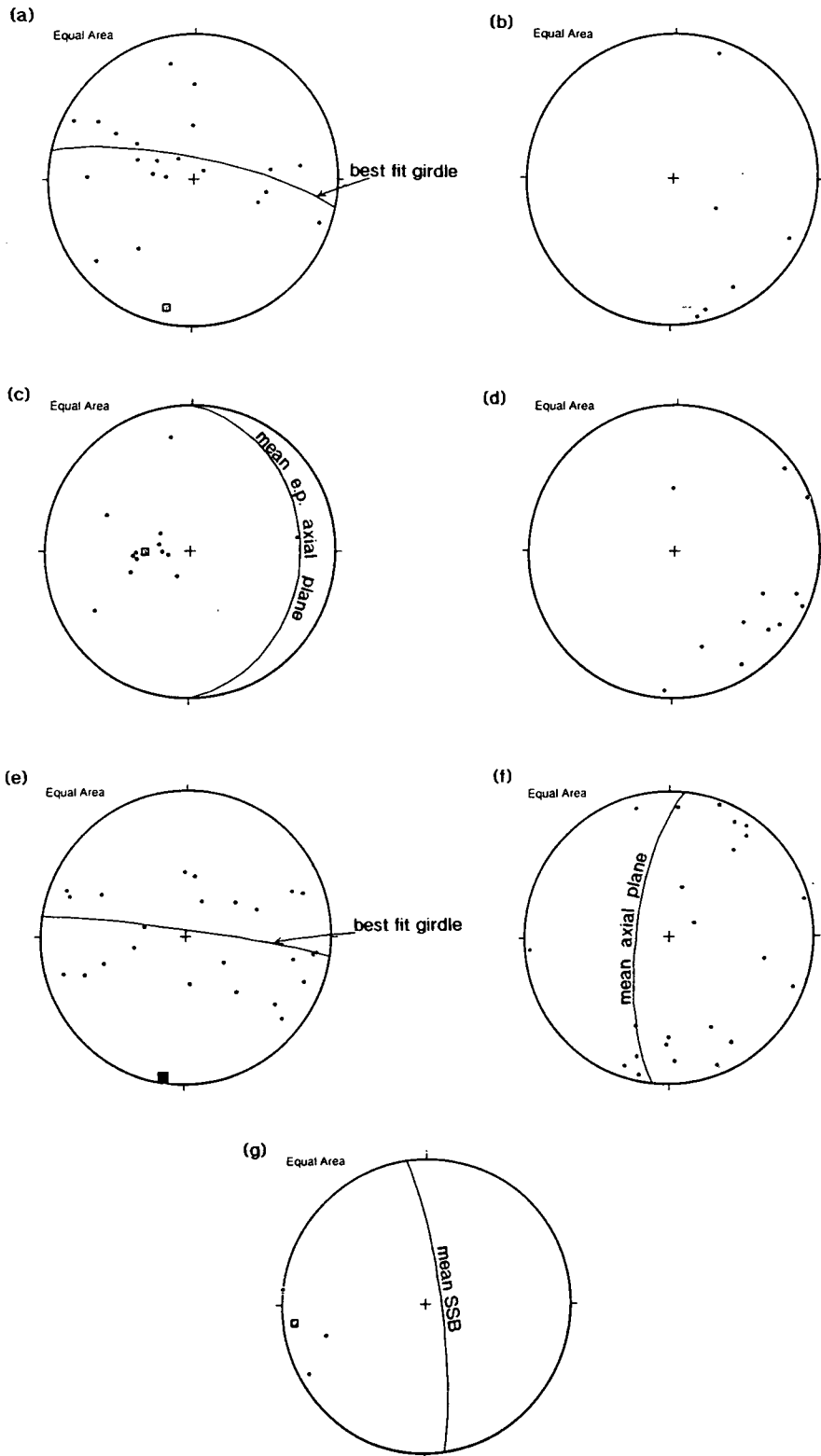


Figure 3.11 Equal area stereonets of structural data from the country rock rafts at Lough Agher. (a) Poles to foliation; (b) mineral lineation; (c) poles to early fold axial planes and crenulation cleavage; (d) early fold axes; (e) poles to main fold axial planes and crenulation cleavage; (f) main fold axes; (g) poles to sinistral shear bands.

Early folds

Most of the data relating to the early folds was collected from a relatively restricted area (see Fig. 3.12), and possibly from the flat limb of a later structure, so the spread of the data is not large. The early structures are mainly tight to isoclinal structures developed on a centimetre- to metre-scale. They refold an early foliation parallel fabric. A moderately strong axial planar crenulation cleavage is associated with the early fold structures. Poles to axial planes and early crenulation cleavage are quite tightly clustered, striking N-S and dipping shallowly to moderately towards the E (see Fig. 3.11c). In the field, these folds verge up to an antiform in the east. Most of the early fold axis plunge shallowly to the SE, although individual plunges vary from moderately northwards to shallowly southwards (Fig. 3.11d). This spread of data could be due to curvilinearity through the strike of their axial planes, producing a whaleback geometry, although this was not observed directly in the field. Alternatively, the spread of fold plunge could be due to later refolding, for which there is field evidence.

Main folds

In the field, these are irregularly developed close to tight folds, developed on a centimetre- to metre-scale. A well-developed, spaced crenulation cleavage is developed and displays a convergent fanning pattern around the main phase fold hinges. They refold the earlier fold structures producing Type III hook interference patterns (see Ramsay 1967; Plate 3.8). On average, the axial planes and crenulation cleavage strike NNE-SSW, and have subvertical dips (see Fig. 3.11e). The poles fall on an ESE-WNW girdle; this spread may reflect the observed fanning nature of the axial planes, or may indicate that they are curvilinear, or that raft rotation, or later folding has occurred. There is no field evidence for the latter process. The fold axes generally plunge either NNE or SSW (Fig. 3.11f) with variable plunges consistent with the curvilinearity of the fold hinges as observed in the field. The folds are curvilinear through the strike of their axial planes (i.e. through the horizontal), and as a result, they tend to show double vergence in the field, having dextral (Z) vergence when they plunge S (Plate 3.9), but sinistral (S) vergence when the plunge N and E. All these folds verge up towards an antiform in the west. A well-developed mineral lineation, defined by phyllosilicates and sillimanite, is often preserved in the hinge zones of the main folds, where early fold structures are also preserved; here it is refolded to plunge parallel to the main fold



Plate 3.8 Example of type III (hook) interference patterns (see area immediately below pen) in pelitic country rock rafts at Lough Agher. These are produced in this area due to main phase folds refolding early phase folds.



Plate 3.9 Example of a curvilinear main phase fold in the country rock rafts at Lough Agher. In this example the double vergence can be observed, since where the folds plunge south they exhibit dextral vergence ('Z' in photo) and where they plunge north and east they exhibit sinistral vergence ('S' in photo).



Plate 3.10 Example of an early mineral lineation being folded around the hinge of a main phase fold. The pencil lies approximately parallel to the early mineral lineation.



Plate 3.11 Rare millimetre wide sinistral shear bands with millimetre to centimetre-scale offsets exposed in hornfelsed country rock rafts at Lough Agher. They are filled with quartzofeldspathic granite veins. Sense of shear indicated by white arrows.

axes (see Plate 3.10). In other areas throughout the rafts, however, sillimanite randomly overprints all pre-existing structures, suggesting that its growth post-dates the formation of all folding events, and that in cases where it is aligned parallel to early fabrics, it has mimetically overgrown earlier mineral alignments.

The latest structures seen in this area are rare millimetre- to centimetre-wide sinistral shear bands, with cm-scale offsets, which are filled with quartzo-feldspathic granite veins (see Plate 3.11). These structures are less abundant than in the 'migmatites' which will be discussed in section 3.1.8 of this chapter and in common with the latter structures, they are believed to be related to the main emplacement event (see section 3.2). Where observed in Lough Agher, they generally strike NNW-SSE and have steep to subvertical ENE dips (Fig. 3.11g). They therefore lie anticlockwise of the mean foliation, as would be expected for sinistral shear.

3.1.6.b *Ardveen*

Figure 3.12 shows the position of Ardveen relative to Lough Agher and Meencorwick and also illustrates that the country rock xenoliths at Ardveen are much smaller than those at Lough Agher. Figure 3.13 (reproduced from Scott 1974) illustrates the fact that the country rock rafts have an approximate N-S orientation; it is the view of this author that this preferred orientation of country rock xenoliths is, in part, a product of rotation and reorientation of the xenoliths in relation to granite emplacement, since the structural data from this area show a large scatter.

The lithology of the xenoliths is dominantly semipelite, with subordinate calcareous and psammitic horizons. The metamorphic mineral assemblage is dominated by fibrolitic sillimanite after andalusite, with variable amounts of biotite, garnet and cordierite. Semipelitic rafts tend to be sub-rounded, perhaps indicating that there has been some assimilation of this lithology, possibly at depth (Scott 1974). Contacts with more psammitic xenoliths tend to be more irregular and angular due to their more competent nature. Semipelitic horizons exhibit a well-developed compositional banding, which forms the dominant foliation in the rafts of the Ardveen area. The data exhibits a very wide spread (Fig. 3.14a), probably as a result of syn-emplacement rotation of individual rafts about a strike-parallel, NNW

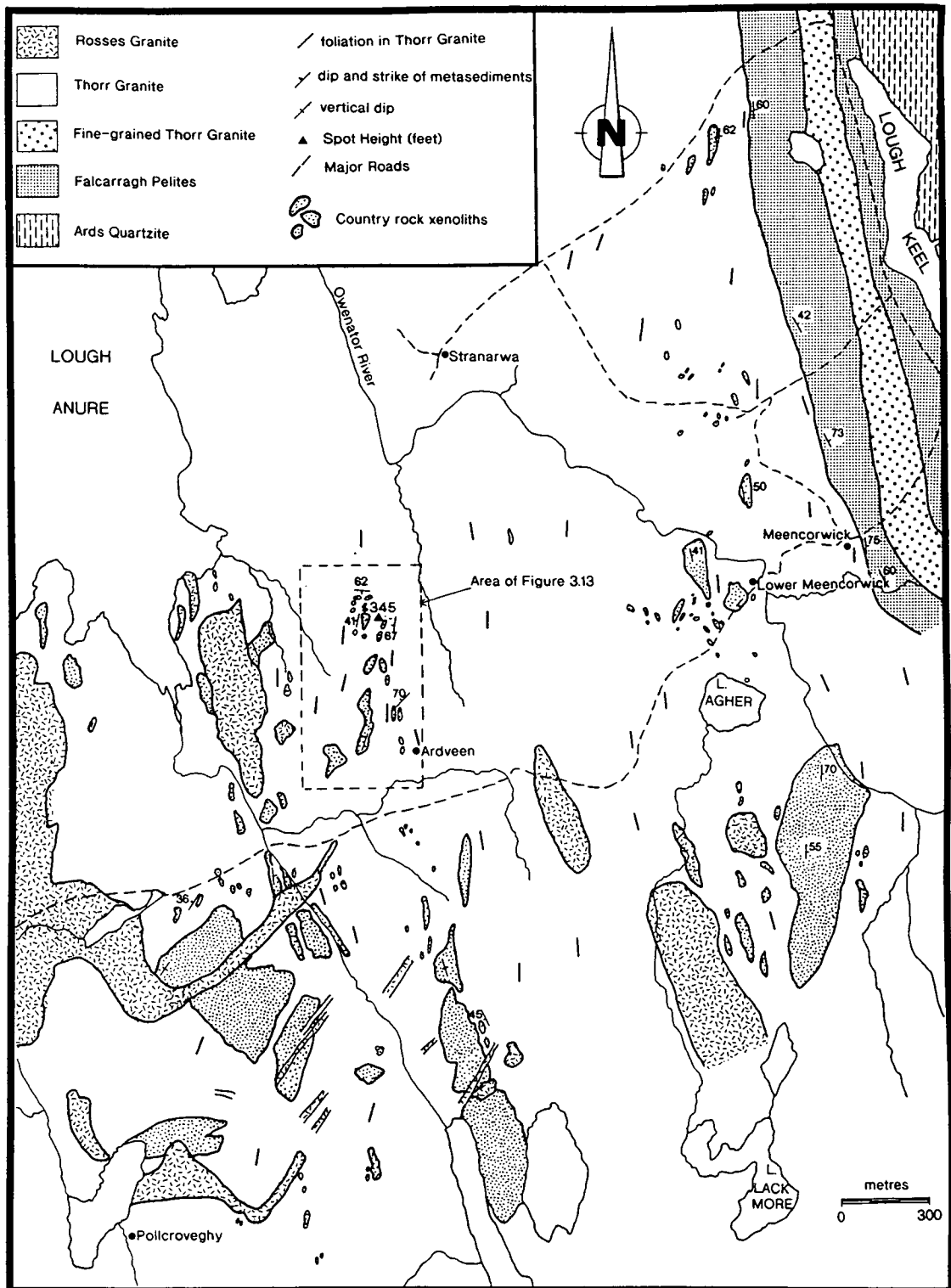


Figure 3.12 Sketch map of the area around Thorr, Meencorwick, Lough Agher and Ardveen showing position of granite contact and distribution of country rock rafts.

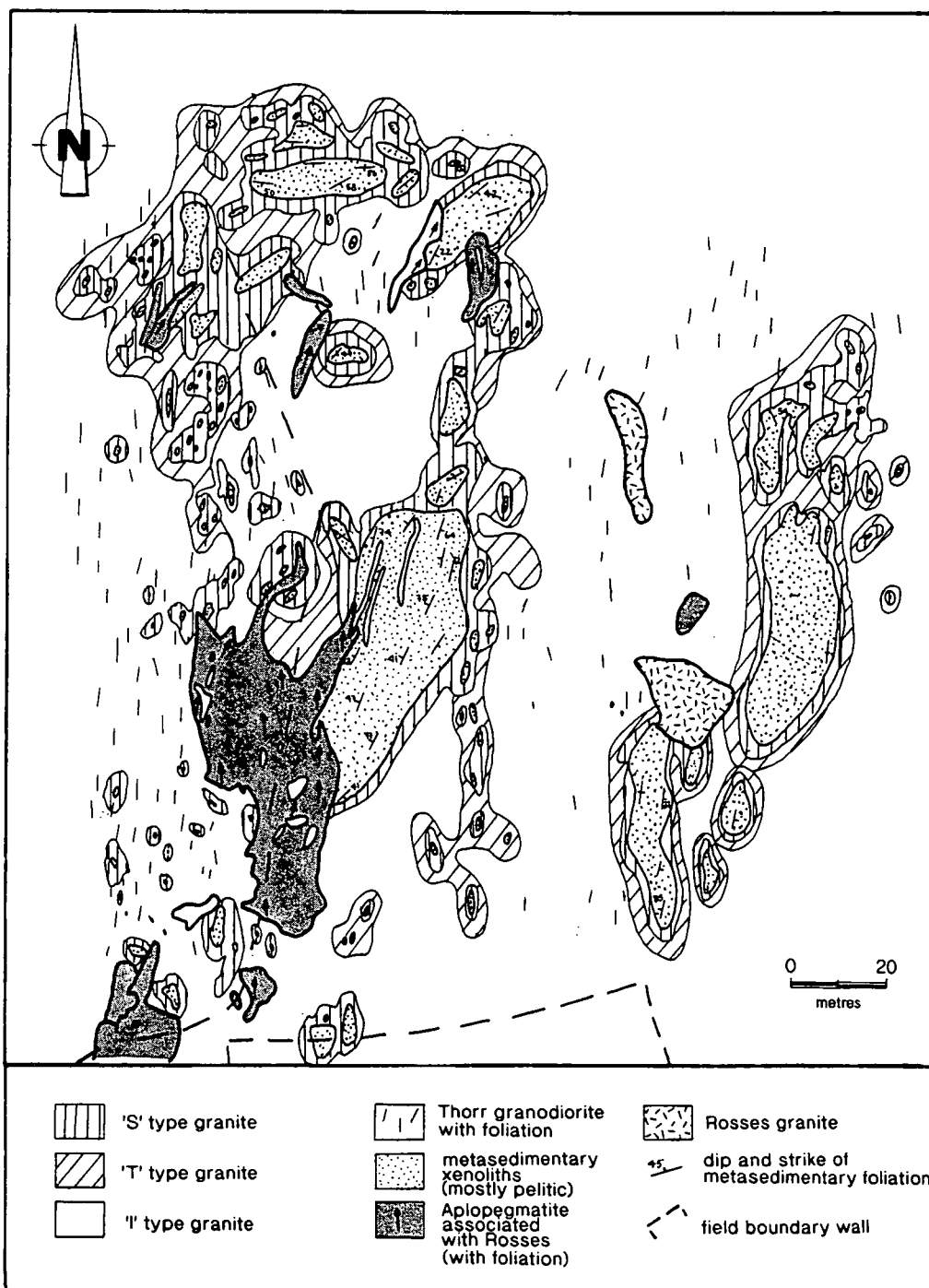


Figure 3.13 Detailed sketch map of the area around Ardeen showing the position of country rock rafts (after Scott 1974).

ARDVEEN

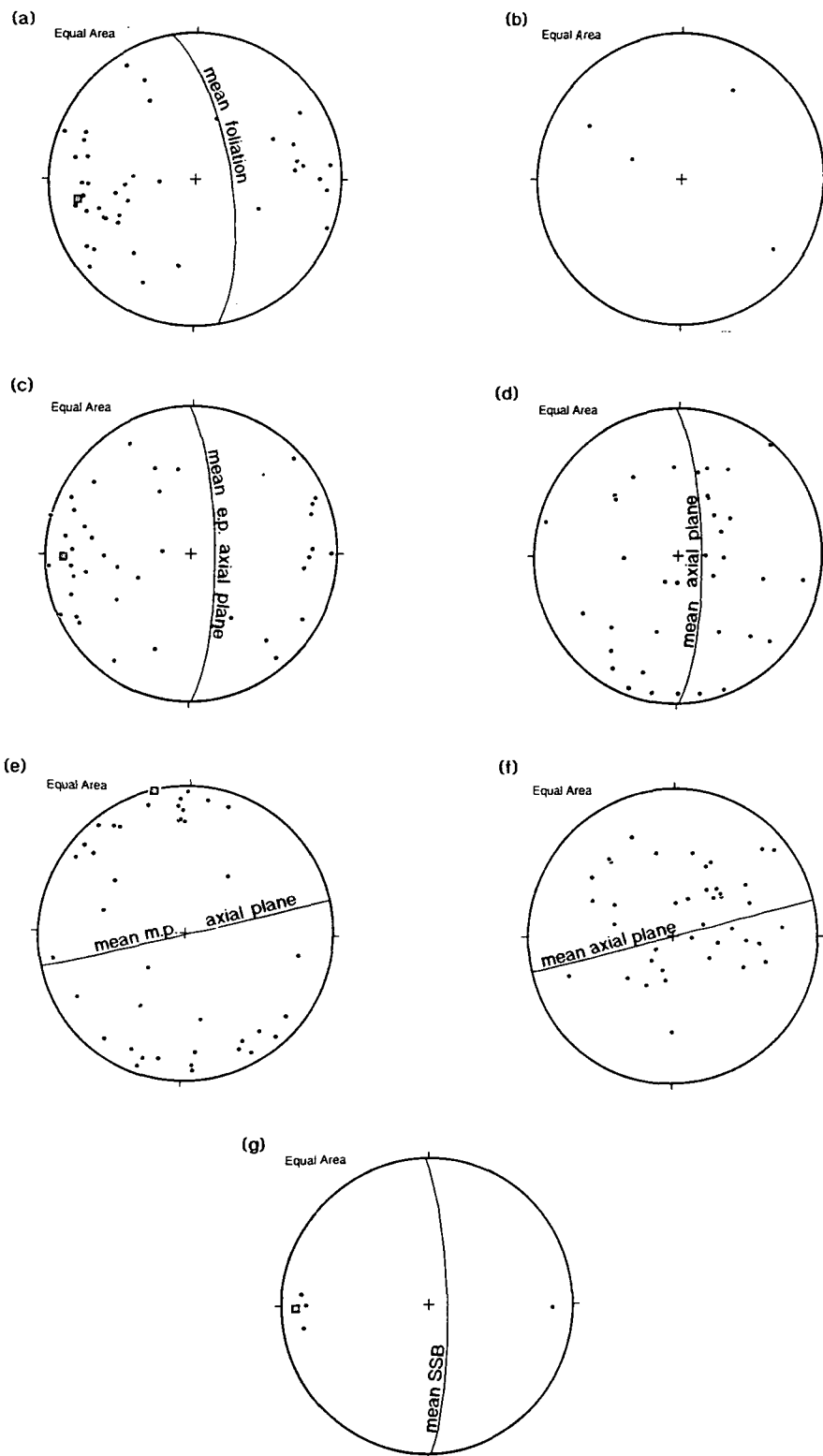


Figure 3.14 Equal area stereonets of structural data from the country rock rafts at Ardveen. (a) poles to foliation; (b) stretching lineations; (c) poles to early fold axial planes and crenulation cleavage; (d) early fold axes; (e) poles to main fold axial planes and crenulation cleavage; (f) main fold axes; (g) poles to sinistral shear bands.

axis, the mean foliation plane striking NNW and dipping steeply towards the ENE. A stretching lineation is infrequently observed, but where seen it is an alignment of phyllosilicate and sometimes sillimanite minerals with a very random plunge (Fig. 3.14b), possibly due to a combination of refolding and raft rotation.

The rafts here are similar to those at Lough Agher, in that there are two fold phases; early and main fold structures.

Early folds

These are not often well preserved, but a strong axial planar crenulation cleavage and bedding-cleavage intersection lineation can be recognised in zones of lower strain within the flat limbs and sometimes the hinge zones of the main fold structures. Where the early folds are exposed, they are centimetre- to metre-scale, tight to isoclinal folds, which deform an earlier fabric parallel to the compositional banding. Cleavage and intersection lineation data are widely spread (Fig. 3.14 c&d), probably due to a combination of refolding by the main phase structures and raft rotation during granite intrusion. The axial planar cleavage shows a weak clustering about a N-S striking plane, with the average plane dipping very steeply towards the E (Fig. 3.14c). The plunge of the intersection lineation (Fig. 3.14d) shows no clear trend. No conclusive vergence data on the early folds was observed in the field.

Main folds

These are generally close to tight structures, developed on a centimetre- to metre-scale, similar to the main phase folds observed at Lough Agher. An extremely well-developed axial planar crenulation cleavage is associated with this set of folds. Metamorphic index minerals, particularly biotite and sillimanite often grow mimetically along the cleavage planes and are also seen to randomly overprint the axial planar fabric, demonstrating that the growth of these metamorphic minerals post-dates the development of the main folds. Data relating to the axial planes and fold axes of the main folds is plotted in figure 3.14e & f respectively. In common with the other structural data from the Ardveena area, these data sets have a wide spread, though in this case the spread can only be accounted for in terms of syn-granite emplacement rotation of country rock xenoliths, since no folds later than this main phase were observed in the area. A crude clustering of poles to axial planes can be recognised in Fig. 3.14e,

indicating a subvertical, ENE-WSW striking mean axial plane, however, an average fold plunge is indeterminable. No conclusive vergence data relating to these folds was observed in the field, but the mean axial plane calculated on the stereonet lies clockwise of the mean foliation plane which, assuming a mean NE fold plunge, might indicate that the main folds verge up to an antiform in the west. However, without observational confirmation this interpretation can only be tentatively suggested.

As at Lough Agher, the only structures overprinting the main phase folds are rare mm- to cm-wide sinistral shear bands, with cm-scale offsets, that are filled with quartzo-feldspathic granite veins, that here strike N-S and are subvertical (see Fig. 3.14g). This average orientation is clockwise of the mean foliation plane. However, there is a wide spread of foliation data and only a few shear bands, and where the shear bands are observed, they lie anticlockwise of the foliation.

3.1.7 Crollly - Thorr - Meencorwick

This area, in addition to the raft trains at Lough Agher and Meencorwick, has been discussed by several workers in the past fifty years, including Gindy (1953b), Mercy (1963) and Pitcher & Berger (1972). For the purpose of this thesis, the author has subdivided the description of the so-called migmatites (Pitcher 1953; Pitcher & Berger 1972) into two sections. Firstly, the structure of the country rocks east of the main contact and secondly, the nature of the migmatite - quartzite country rock contact, as exposed south of Lough Keel (see Map 1).

3.1.7.a Structure of the country rocks east of the main granite - country rock contact

A zone of variably exposed, 'migmatized' (Pitcher 1953; Pitcher & Berger 1972) country rocks was traced from Crollly southwards through Meencorwick, past Tor School to Lough Ascardan (see Map 1). The dominant lithology is hornfelsed semipelite, with subordinate amounts of psammite/quartzite and calc-pelite. Metamorphic minerals are abundant as a result of contact metamorphism. Sillimanite (after andalusite) is dominant, but cordierite and biotite are also abundant, whilst garnet is variably developed according to the lithology, being most common in pelite and calc-pelite horizons.

LOUGH KEEL - THORR - MEENCORWICK

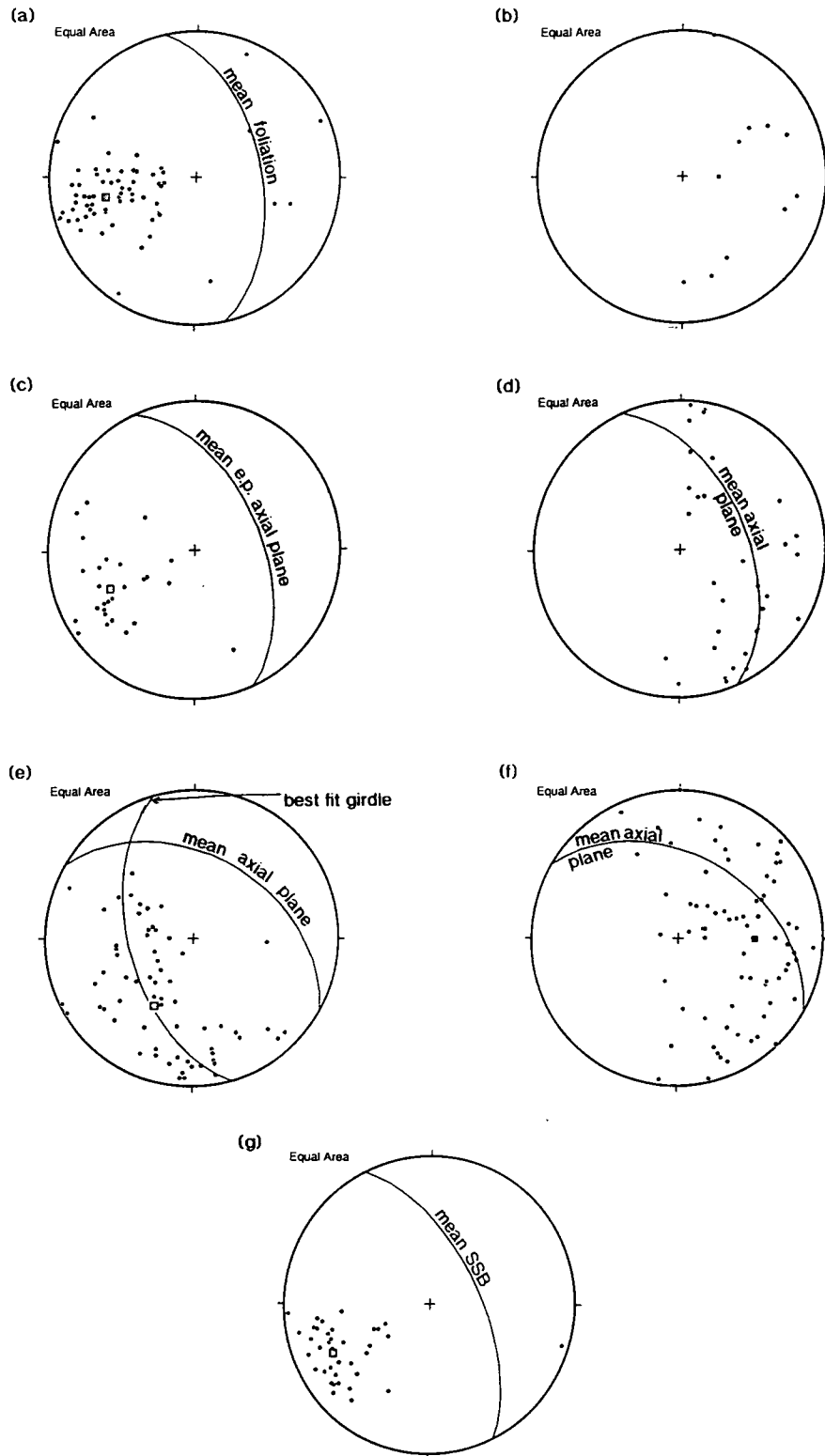


Figure 3.15 Equal area stereonets of structural data from the mobilised country rocks at Thorr. (a) poles to foliation; (b) stretching lineation; (c) poles to early fold axes and crenulation cleavage; (d) early fold axes; (e) poles to main fold axial planes; (f) main fold axes; (g) poles to sinistral shear bands.

Structurally, these rocks are apparently more complex than the other country rocks surrounding the Thorr pluton. This is reflected by their chaotic appearance that seems to have arisen due to deformation synchronous with mobilisation in response to rapid heating by the granite (see below). All through the section, the chaotic folding is overprinted by uniformly striking sinistral shear bands filled with quartzo-feldspathic granite sheets that become increasingly common towards the granite contact. Earlier deformation structures pre-dating the chaotic folds are mostly transposed into parallelism with the regional foliation.

These rocks have a well-developed foliation, that is best described as compositional, or even segregation banding. On the stereonet (Fig. 3.15a), poles to the foliation show a distinct clustering in the WSW sector of the net, indicating that the mean foliation plane dips moderately to steeply towards the ENE and strikes NNW-SSE. The data show a slight spread along an ENE-WSW girdle. A mineral lineation, defined by phyllosilicates and occasionally sillimanite, can sometimes be recognised along the foliation planes (Fig. 3.15b). The lineations show a fair spread, possibly due to early development of this lineation in the deformation history and subsequent refolding. In general, however, the plunge of the lineation is shallow to moderate in an easterly direction, although the direction of plunge varies from ENE to S.

As in the country rock rafts, the migmatites contain two sets of fold structures; early folds and a main set of chaotic structures.

Early folds

Plate 3.12 shows a typical early fold structure within the mobilised country rocks. They clearly deform an earlier bedding-parallel fabric, are usually isoclinal and are developed on a variety of scales, from centimetres through to metres. They are associated with a very well-developed, millimetre-spaced crenulation cleavage, which is often almost completely transposed. The axial planes to these folds, and their associated crenulation cleavage, are mainly NW-SE striking, and dip moderately to shallowly towards the NE (Fig. 3.15c). The poles to early fold axial planes show some spread which is interpreted as being a product of fanning of the early fold axial planes and the associated crenulation cleavage about the fold hinges; alternatively it may be due to a curvilinear geometry. The early fold axes have variable

plunges (Fig. 3.15d) towards the east. They are spread along a NNW-SSE trending girdle; this is interpreted as reflecting slight curvilinearity, for which there is field evidence. The vergence of these folds seen in the field indicates that the early folds verge up to an antiform in the W to SW.

Main folds

These are tight to isoclinal, centimetre- to metre-scale folds (Plate 3.13) that are markedly curvilinear through the strike of their axial planes (i.e. through the horizontal), producing 'whaleback' to 'sheath fold' geometries on all scales seen in the field. The apparent vergence (i.e. whether they are 'S' or 'Z' shaped) of these folds in the field can be ambiguous due to the curvilinearity of the fold hinges, which results in the double vergence patterns characteristic of sheath folds. The geometry of these folds is summarised in figure 3.16. The significance of the vergence of these folds will be discussed later in this section. They refold all earlier structures and also fold fibrolite after andalusite and sillimanite porphyroblasts.

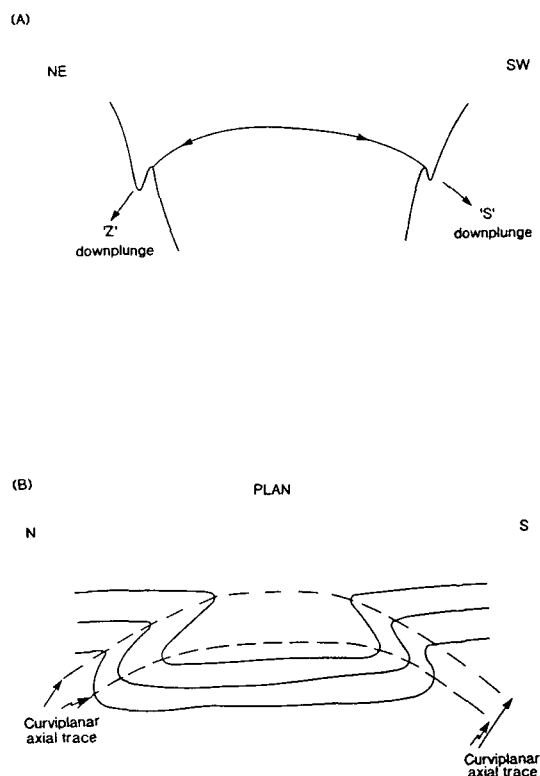


Figure 3.16 Geometry of the main phase folds in the mobilised lithologies. (A) illustrates their geometry in cross-section and (B) illustrates their geometry in plan view.



Plate 3.12 Typical centimetre-scale, isoclinal early fold (centre of plate) in mobilised country rocks near Tor School (see Maps 1 & 2). These deform an earlier bedding-parallel fabric.



Plate 3.13 Typical two-dimensional form of a main phase fold in the mobilised hornfelses exposed from Croll to Thorr. The curvilinear nature of these folds is not often exposed on the scale of a single outcrop.

Stereonet (e) and (f) in figure 3.15 are plots of the poles to axial planes and fold axis plunges respectively. The mean axial plane strikes NW-SE and dips moderately towards the NE. The data are spread along a NW-SE striking girdle, that is interpreted as reflecting a curvilinear arrangement of the axial surfaces, a feature that was observed in the field. The fold axes are strongly curvilinear as illustrated by the wide spread of data. This is widely dispersed in the eastern half of the net, rather than forming a clear girdle. This is interpreted as another reflection of the fact that the axial surfaces are curvilinear (as schematically illustrated in Fig. 3.16b), possibly as a result of lack of cohesion within the migmatites at the time of formation of these folds (see discussion in section 3.2). The stereonet data supports the vergence data observed in the field; these folds consistently verge up towards an antiform in the NE, having 'Z' vergence when northward plunging and 'S' vergence when southward plunging (see Fig. 3.16a).

Like the rafts at Lough Agher and Ardveen, the only structure post-dating the main fold phase in the migmatites are centimetre-wide sinistral shear bands filled with fine-grained granite identical to that found in the marginal facies of the adjacent Thorr pluton. Stereonet (g) in figure 3.15 summarises the shear band data. The poles form a cluster on the stereonet, to which the mean plane has a NE-SW strike and a very steep dip towards the NE. The sinistral shear bands plot anticlockwise of the foliation, as one would expect in a sinistral shear regime. Offsets on these sinistral shears vary here from less than a centimetre up to tens of centimetres (see Plate 3.14). These shear bands appear to be associated with the main phase of emplacement of the Thorr granite and their density increases towards the main granite contact, as the migmatites becoming increasingly mobilised. In addition, there are two large, prominent gullies striking subparallel with the sinistral shear bands in this section (see Map 1). It is difficult to match up markers in the country rocks across these features, leading the author to tentatively propose that they are larger shear zones related to granite emplacement.

On the basis of the observed occurrence and offsets on these sinistral shear bands, it is possible to crudely estimate minimum and maximum offsets associated with these displacements during granite emplacement; possible earlier displacements associated with mobilisation cannot be estimated. In a completely exposed 10m wide zone nearest to the contact with the granite (see Map 1) 50 sinistral shear bands were counted, that is one sinistral shear band every 20cm. Since the sinistral shear bands mostly



Plate 3.14 Commonly occurring centimetre wide sinistral shears overprint all deformation structures in the mobilised country rocks. They commonly exhibit centimetre-scale offsets and are filled with quartzofeldspathic granite veins.

show displacements of between 1cm and 10cm, then upper and lower estimates of displacement can be calculated for zones of specific widths. The lower and upper zone widths for the rough calculations are taken as 100m and 500m respectively. These values were chosen because the zone of prolific sinistral shear bands was estimated as being approximately 100m wide in the field and the whole migmatite zone was estimated as approximately 500m wide (see Map 2). The rough calculations are outlined below:

(A) For a 100m wide zone:

(I) *Lower limit*

- if each sinistral shear band has approximately 1cm displacement,
- and if the mobilised zone is 100m wide, the total offset will be:

(Width of shear zone / 20) x displacement on each shear band

$$= (10\ 000 / 20) \times 1$$

$$= 500\text{cm} = 5 \text{ metres}$$

(II) *Upper limit*

- if each sinistral shear band has approximately 10cm offset
- and if the mobilised zone is 100m wide, the total offset will be

$$= (10\ 000 / 20) \times 10$$

$$= 5000\text{cm} = 50 \text{ metres}$$

(B) For a 500m wide zone:

if one assumes from the above a maximum of, say 20cm spacing of sinistral shear bands, and the whole migmatite zone is estimated to be approximately 500m wide, then there is an average of 5 sinistral shear bands per metre. So,

(I) *Lower limit*

- if each sinistral shear band has approximately 1cm displacement,
- and the migmatite zone is 500m wide, then the total offset

$$= (50\ 000 / 20) \times 1$$

$$= 2500\text{cm} = 25 \text{ metres}$$

(II) *Upper limit*

- if each sinistral shear band has 10cm offset
- if the migmatite zone is 500m wide, then the total offset

$$\begin{aligned} &= (50\,000 / 20) \times 10 \\ &= 25000\text{cm} = 250 \text{ metres} \end{aligned}$$

These calculations are obviously very rough estimates, and true displacement may be significantly different. For example, if the prominent gullies mentioned above are larger shear zones, then much larger sinistral movements may have occurred in addition to those recorded by the centimetre-scale shear bands. On the other hand, displacements may be smaller as the spacing of the shear bands is significantly less than one every 20cm in the outer parts of the shear zone, away from the main granite margin.

3.1.7.b *The nature of the migmatite - quartzite country rock contact as exposed south of Lough Keel*

Disagreement and confusion have arisen in the literature concerning the stratigraphic and large-scale structural significance of the contact between the quartzites and pelites in the region south of Lough Keel (see section 3.2.7). As discussed below, there is good evidence that the boundary has suffered intense strain prior to granite intrusion, but this region also preserves evidence of syn-intrusion sinistral shear.

The deformation chronology in a small (c. 500m) section south of Lough Keel (Fig. 3.19 & Map 1) is extremely complicated, with at least three lineations (some down-dip and some strike-slip; c.f. Rickard (1963)) occurring in outcrops of quartzite, pelite, semipelite and calc-pelite. The intensity and orientation of the lineations varies according to lithology. Down-dip lineations mostly occur in the quartzites and probably relate to movement on the tectonic slide and the early development of the Crockator syncline and Thorr Bend (see Ch. 2). Strike-slip lineations mostly occur in the pelite horizons and relate to mobilisation and granite emplacement.

Plate 3.15 is a general view of the contact between the quartzite and the pelite. The quartzite generally dips shallowly to moderately towards the ESE, whereas the pelites are more steeply dipping. The



Plate 3.15 General view, looking east, of the tectonic contact between quartzite and pelite country rocks south of Lough Keel.



Plate 3.16 Well-developed down-dip mineral lineation in quartzite above the tectonic slide. This lineation has a typical rodded appearance and it plunges moderately towards the ESE and SE down the dip of the foliation.

strain throughout this small section is very high, and this is particularly obvious in the quartzites, since remnants of folds can often be seen that are now completely transposed parallel to the main foliation. The relatively high strain within the quartzite unit gives it a platy or flaggy appearance in the field. Thin sections of the quartzite from this area have an annealed texture, typical of a deformed rock that has then been baked within a contact aureole with large quartz grains entirely enclosing aligned phyllosilicate minerals. There is some evidence of post-annealing subgraining, but this is a relatively minor feature. Mineral lineations within the quartzite typically plunge moderately down the dip of the foliation towards the ESE-SE and have a rodded appearance (see Plate 3.16). However, within discordant high strain zones in the quartzite, this down-dip lineation is rotated into a much shallower NNE - NE plunge (see Plate 3.17). These discordant shear zones have the form of metre-scale sinistral bands (see Plate 3.18). Furthermore, shallowly plunging to strike-parallel lineations are common throughout the various pelitic bands that occur interbanded with the quartzite and below the tectonic slide. These semipelitic and calc-pelites are also full of sinistral shear criteria, including sinistral shear bands (Plate 3.19), giving a top to the north sense of shear. This suggests that the pelite horizons have acted as foliation-parallel sinistral shear zones. Thin (1-2cm) granite veins are also seen along the section. They are often disrupted and augened, again with a top to the north sense of shear, and typically carry solid state deformation fabrics. Early boudins within the quartzite get refolded/rotated within the section. Quartz veins within the boudin necks display apparent dextral offsets (see Plate 3.20). This feature could also be consistent with an overall sinistral shear couple in a bookshelf type model, as outlined in figure 3.17. The significance of these structures in terms of granite emplacement will be discussed in section 3.3 of this chapter.

It is suggested that the pelites have accommodated substantial amounts of sinistral shear due to rheological softening associated with mobilisation and granite emplacement. These strains overprint the regional down dip lineations preserved in the less reactive, more resistant quartzites. In addition, there is evidence for high pore fluid pressures synchronous with granite emplacement within the quartzites as intrusion breccias occur in places (see Plate 3.21). The brittle behaviour of the quartzite contrasts strongly with the ductile response of the pelites at this time and it is this difference that accounts for much of the complexity in this region.



Plate 3.17 Shallowly inclined mineral lineation associated with a high strain zone exposed in quartzite along the tectonic contact south of Lough Keel. This lineation is defined by an alignment of phyllosilicate minerals and quartz. It generally plunges gently towards the north along the strike of the foliation.



Plate 3.18 Examples of high strain zones within the quartzites exposed south of Lough Keel. In this case the high strain zones have the form of metre-scale sinistral shear bands.



Plate 3.19 Examples of sinistral shear bands and within mobilised pelitic lithologies immediately beneath the tectonic slide exposed south of Lough Keel. Such sinistral shear criteria are believed to have developed in association with granite emplacement.



Plate 3.20 Boudinage structures in quartzite adjacent to the tectonic slide exposed south of Lough Keel. Boudin necks are filled with quartz veins. Rotation of these structures into their present orientation is believed to have been facilitated by dextral sliding along quartz veins in boudin necks in response to an overall sinistral shear couple.



Plate 3.21 Example of an intrusion breccia within semi-pelites adjacent to the tectonic high strain zone exposed south of Lough Keel. Breccia consists of clasts of semi-pelite set in a fine- to medium-grained quartzo-feldspathic matrix.



Plate 3.22 Photomicrograph illustrating that the lithologies beneath the tectonic slide exposed south of Lough Keel are calcareous. Specimen is a diopside - actinolite - epidote hornfels. (Field of view 4.4mm).

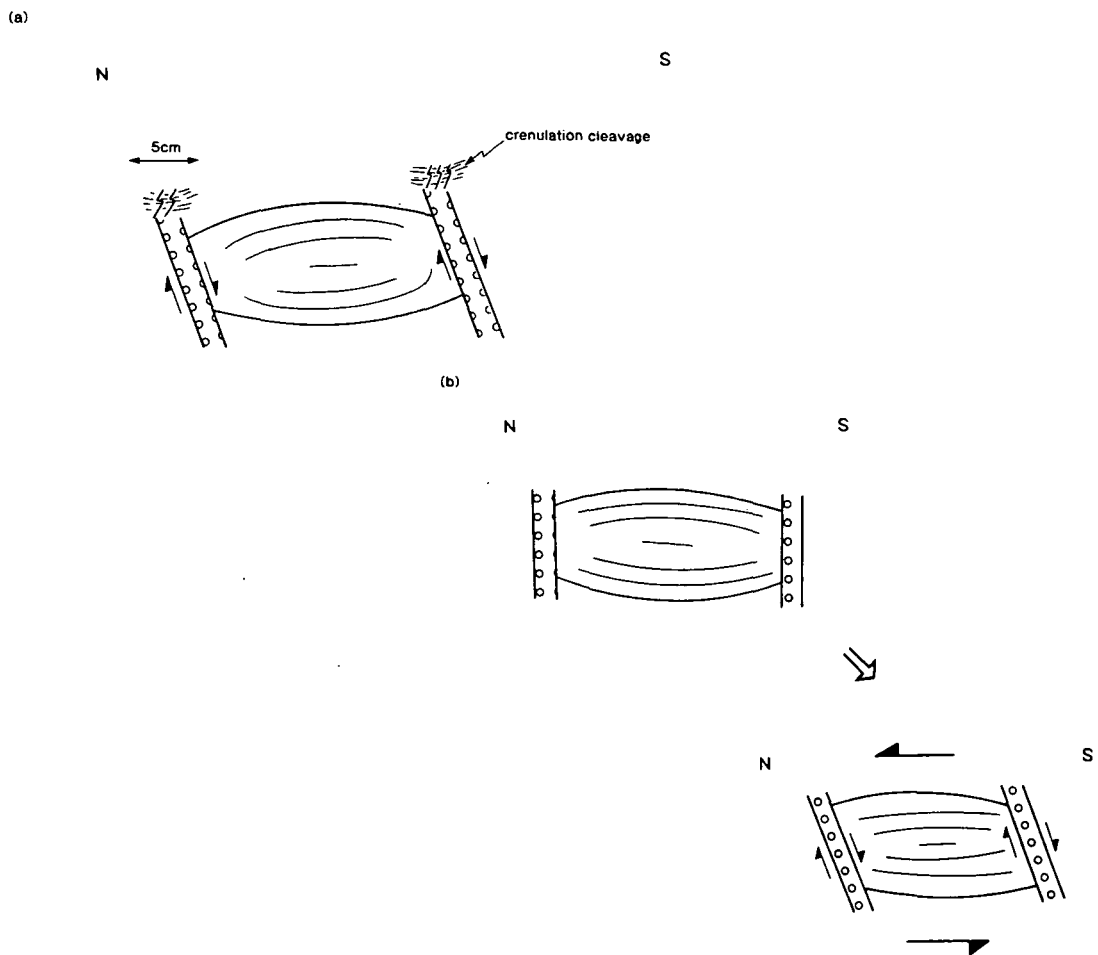


Figure 3.17 Rotated boudin structures found in Ards Quartzite within the high strain zone south of Lough Keel. (a) Sketch of boudins showing the quartz veins in the neck of the structure; (b) bookshelf sliding model for rotation of structures under conditions of sinistral shear (after Hanmer 1986).

3.2 Stratigraphical correlation of the envelope rocks and the xenolithic rafts within the pluton

The stratigraphic units surrounding the Thorr pluton are shown in Map 2, together with the broad stratigraphic affinities of the country rock xenoliths within the granite. It is not an aim of this thesis to produce a detailed stratigraphy for this area, but, it is necessary to have an understanding of the distribution of the lithologies, in addition to their deformation history, since these factors may have a direct influence on the mode of emplacement of the pluton. The stratigraphy is briefly summarised in the following sections.

3.2.1 *Binanea Strand, Tory Island, Tievelehid and Bloody Foreland Mountain*

In these areas, the granite is seen in contact with quartzite, correlated with the Ards Quartzite (see Fig. 2.2; Pitcher & Berger 1972 and references therein). Ards quartzite forms a continuous strip along the east of the granite on the mainland from Binanea Strand to Crockator. At the Bloody Foreland Mountain, quartzite forms a cap that sits on top of the granite. The strike and dip of the bedding in the quartzite at the Bloody Foreland Mountain is not parallel to that in the quartzite at Binanea Strand. This may indicate that the quartzite cap was rotated at the time of granite intrusion, or, it may equally be attributed to earlier large-scale folding within the quartzite. Such folding can be observed from changes in vergence and direction of dip around the Tievelehid area (see map accompanying Pitcher & Berger 1972). Large xenoliths of quartzite, up to 20m wide, are found within the granite along the north coast, particularly around the area of the Bloody Foreland (see Map 2). Pitcher and Berger (1972) correlated the quartzite xenoliths within the granite on the north coast with the Ards quartzite. The present author can find no evidence to the contrary.

3.2.2 *Curran's Point and Inishbofin Island*

Country rocks seen at these localities are dominantly pelites and semipelites that have been variably metamorphosed within the aureole of the Thorr Granite. The pelitic lithologies at Curran's Point have been correlated with the Upper Falcarragh Pelites (Pitcher & Berger 1972). The lithologies present on Inishbofin Island were described by Rickard (1962), who suggested that the pelites in the southeast of the island could be correlated with the Upper Falcarragh Pelites. However, he also suggested that they are conformably overlain to the west by semipelites that become more quartzose towards the west and preserve small-scale sedimentary structures, such as 'candle-flame' and slump structures. He called these rocks the Inishbofin Banded Semipelites and correlated them with the Loughros Formation (see Fig. 2.2) on the basis that they have a higher quartz content than the Falcarragh Pelites and preserve sedimentary structures. The present author found no evidence of sedimentary structures in the pelites and semipelites at Curran's Point and on Inishbofin Island and is therefore unable to comment on whether the semipelitic

lithologies to the west of Inishbofin are part of a younger formation than those on the mainland, or whether they simply represent the upper part of the Upper Falcarragh Pelites. In either case the strata appear to be right way up.

3.2.3 Aranmore Island

The dominant country rock lithology on this island is Ards Quartzite (Pitcher & Berger 1972). This author, however, found the granite in contact with pelite and semipelite (see Map 2 NC 682151 & 679149). The only pelitic lithology recorded from the surrounding region is a small outcrop of Ards Pelite on Calf Island, to the east of Aranmore Island (see Map 2). The pelitic rocks found on Aranmore lie on an approximately straight line trending SSW from Calf Island and parallel to the granite country rock contact (see Map 2), so it seems fair to correlate these units. This relationship is consistent with granite emplacement parallel to the pre-existing lithological boundary that is now almost completely obscured by the Thorr Granite.

3.2.4 Maghery

Inverted Ards Quartzite and right way up Ards Pelites have been described in this area by Pitcher & Berger (1972) and Meneilly (1982). The Ards Quartzite is intruded by pre-granite dolerite sills that have been subjected to low grade regional metamorphism. The granite is seen in contact with Ards Pelites at the coast around Maghery, but inland towards the SE, it is in contact with Ards Quartzite. Most of the country rock xenoliths within the granite around the Maghery - Lough Illion - Toberkeen area are pelitic and semipelitic (see Map 2), and their strike is roughly concordant with the strike of the foliation in the envelope rocks in this area, suggesting that the rafts form a ghost stratigraphy (c.f. Pitcher & Berger 1972).

3.2.5 Lettermacaward, Lough Agher and Ardveen

The country rock stratigraphy in this area is dominated by Falcarragh Pelites (Upper and Lower) and Falcarragh Limestones (Pitcher & Berger 1972 and references therein; Meneilly 1982). The Thorr

granite is seen in contact with Upper Falcarragh Pelites to the SE towards the Gweebarra River, and Falcarragh Limestones further inland towards the NW (see Map 2). Country rock rafts within the granite are dominantly hornfelsed Upper Falcarragh Pelites (Pitcher & Berger 1972), including those at Lough Agher and Ardveen.

3.2.6 Crolly, Tor School, Meencorwick, Thorr

Considerable debate has arisen concerning the stratigraphic affinities of the pelitic rocks in the Thorr area and the significance of their boundary with the overlying Ards Quartzite. Pitcher (1953) and Pitcher & Berger (1972) referred to the pelitic rocks south of Lough Keel as the Thorr pelitic series. They suggested that they were older than the overlying quartzite, correlating them with the Ards Pelite (see Fig. 2.2). Such a relationship implies that the contact between the quartzite and pelite in this area is stratigraphic, albeit possibly modified by high strain (Fig. 3.18a). In 1959, however, Pitcher & Read correlated the so-called Thorr Pelite with the Cleengort Pelites of the Maas Succession (\equiv Upper Falcarragh Pelites; Fig. 2.2), rocks that are *younger* than the Ards Quartzite. This relationship implies that the contact between the quartzite and pelite is a tectonic boundary, probably a D_2 slide (see Table 2.1), emplacing older Ards Quartzite over younger Upper Falcarragh Pelite (see Fig. 3.18b). Pitcher & Read (1960) and Rickard (1962), however, correlated the Thorr Pelite with a pelite found at the base of the much older Creeshlough Formation (see Fig. 2.2), suggesting that the slide juxtaposed younger Ards Quartzite over these rocks, whilst a thin slice of intervening "black Ards Pelite" is trapped in between these two units (see Fig. 3.18c). The present author can find no evidence to support the suggestions of Pitcher & Read (1960) and Rickard (1962) since samples of pelite collected from beneath the tectonic slide are calcareous, diopside - actinolite - epidote-bearing lithologies (see Plate 3.22), that look similar in hand specimen and thin section to other calcareous horizons found in the rafts of Upper Falcarragh Pelite included within the granite elsewhere in the SE of the pluton. Hence, it is the opinion of the present author that the pelites in this area can be correlated with the Upper Falcarragh Pelites, as suggested by Pitcher & Read (1959). This interpretation implies that these rocks are overlain by the Ards Quartzite due to the presence of the tectonic slide in this area, as outlined in figure 3.18b.

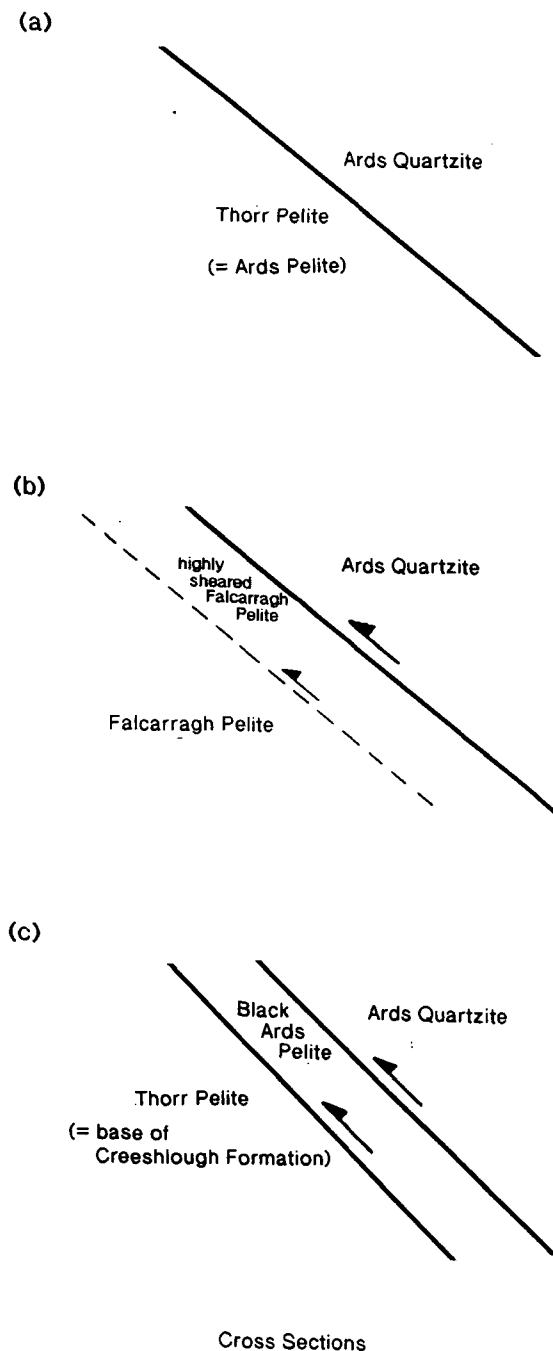


Figure 3.18 Sketches to accompany possible explanations for the juxtaposition of quartzite and pelite south of Lough Keel. (a) Stratigraphical contact between Thorr Pelites (older) and Ards Quartzite (younger) [after Pitcher 1953]; (b) tectonic slide emplaces Ards Quartzite (older) over Falcarragh Pelite (younger) [after Pitcher & Read 1959]; (c) Tectonic slide emplaces Ards Quartzite (younger) over Ards Pelite (older), which is then tectonically emplaced over Thorr Pelite (oldest) [after Pitcher & Read 1960; Rickard 1962].

3.3 Discussion of the structural evolution of the Thorr area, NW Donegal

In most subareas it is relatively simple to identify structures that pre-date granite intrusion and others that are synchronous with emplacement. It is also possible in most cases to correlate structures from one subarea to another, especially folds, on the basis of the geometry of the structures and the type and intensity of the fabrics associated with the deformation. Structures will be correlated in terms of pre- and syn-intrusion across the pluton in the following discussion.

3.3.1 Pre-emplacement deformation structures

3.3.1.a The unmobilised country rocks

There is a strong contrast in the complexity and intensity of strains observed in quartzite country rocks and pelitic (including semipelites and calc-pelites) lithologies. These differences often culminate in the development of high strain zones along the boundaries between pelites and quartzites, such as that seen at Thorr, south of Lough Keel (see Map 1). This is typical of deformation observed elsewhere in the Dalradian (Hutton 1982a).

In most of the subareas where the country rocks are dominantly pelites and semipelites two sets of folds are observed. The earliest of these folds deform bedding-parallel fabrics and are associated with the strong development of an axial planar crenulation cleavage and sometimes, a stretching lineation defined by the alignment of phyllosilicate minerals. All of these early structures are refolded by the main phase folds, which have an associated, but variably developed crenulation cleavage associated with them. Thus the early structures clearly post-date the D_1 event, since they refold the bedding-parallel fabric formed during it; they are, therefore F_2 folds. The main phase folds seen in most subareas must, therefore, be F_3 folds. This means that the fold at Maghera is an F_3 fold, the hinge zone of which is passively intruded by the Thorr Granite. Likewise, the Thorr Anticline, the ghost anticline within the Thorr Granite (see Fig. 2.4), must be an F_3 fold, since it refolds earlier folds. This means that the main phase folds seen in the rafts are parasitic structures that verge up to the Thorr Antiform to the west of Lough Agher and Ardveen (see Map 2). Pitcher & Berger (1972) called this fold a composite F_2/F_3 fold since

they were unable to distinguish F_2 and F_3 structures. It is the belief of this author that the Thorr Anticline cannot be an F_2 structure, since vergence data from early folds within the country rock rafts in the SE of the pluton (see section 3.1) are not consistent with the presence of an antiform to the west of Lough Agher and Ardveen, in the position of the present-day trace of the Thorr Anticline.

On Inishbofin Island only one obvious set of fold structures can be seen. These folds appear to be identical to the main phase folds seen at Curran's Point, nearby on the mainland (see Map 2) in terms of their geometry and the variably developed nature of the axial planar crenulation cleavage. The main phase folds at Curran's Point are F_3 according to the chronology outlined above, since they fold an earlier crenulation fabric and mineral lineation. Likewise, at Lettermacaward, there is only one obvious set of folds, which have a moderately developed crenulation cleavage associated with them. These folds and the crenulation fabric seen with them are similar to the main phase folds at Maghery. Meneilly (1982) described the folds in the Lettermacaward area, paying particular attention to the timing of the Maas syncline and the Mulnamin anticline. Pitcher & Berger (1972) proposed that the Mulnamin anticline is an F_4 structure, refolding the Maas syncline, which they believed to be an F_3 structure. Meneilly (1982), however, could find no evidence for the Mulnamin anticline being younger than the Maas syncline, since the former does not refold more fabrics than the latter. Moreover, hand specimens and thin sections from the two fold structures are almost identical. Meneilly, therefore suggested that these are both F_3 structures. The present author agrees with the interpretation of Meneilly (1982) since the folds seen at Lettermacaward, which verge up to a recumbent anticline in the south, refold an earlier fabric and mineral lineation, similar to the refolded stretching lineation seen at Maghery, which is believed to be of D_2 age.

The origin and timing of the strike swing in the Dalradian country rocks of Donegal has often been related to the emplacement of the Thorr granite. Rickard (1963) proposed that it post-dated the formation of the F_3 folds and that it was, at least partly, the product of forceful intrusion of the Thorr Pluton, which resulted in bending of the pre-existing structures. He cited the example of the Crockator syncline being rotated around the Thorr bend (see Figs. 2.7 & 3.19). However, it is the opinion of this author that the strike swing was achieved before, or at most during the very early stages of emplacement of the Thorr Granite, since the country rock rafts within the granite trend approximately N-S. There is no

evidence of granite emplacement having been forceful in the regions where rafts occur as most granite sheets in the Lough Agher and Ardveen areas intrude rather passively along either foliation planes or fold axial planes. Furthermore, one would expect to find very strong deformation fabrics within the granite itself if its intrusion was associated with forceful emplacement (c.f. Ardara, Holder 1979) and these fabrics should form a continuum through the crystallisation interval, from magmatic state through to solid state and even mylonitic fabrics. However, these phenomena are not present in this area. Fabrics within the Thorr Granite in this area are almost exclusively weakly to moderately developed magmatic state fabrics, that are roughly N-S to NNW-SSE trending, except in the area closest to the Main Donegal Granite, where the magmatic state deformation fabrics are rotated into a WNW orientation and ENE trending solid state fabrics, related to the emplacement of the latter, overprint the magmatic state fabrics (see Ch. 4). In the absence of any strong evidence, such as that outlined above, the present author suggests that the strike swing, at least in this part of Donegal, had occurred before the emplacement of the Thorr Granite.

3.3.1.b *The mobilised country rocks*

The main phase folds within the mobilised pelites are much more chaotic in their geometry and development and appear to be the product of more complex strains than those associated with the main phase folds seen in the country rock rafts at Lough Agher and Ardveen. In addition, their vergence patterns are not consistent with the vergence of the main phase folds in the country rocks rafts at Lough Agher and Ardveen, since they verge up to an antiform in the NE and not the W, yet there is no evidence to suggest the presence of a synform between Lough Agher and Meencorwick. A possible explanation for the discrepancy in vergence outlined above is that the main phase folds seen in the mobilised pelites and semipelites are post- F_3 structures and that have only developed locally in association with mobilisation of the pelites as a consequence of the early stages of granite emplacement. These localised folds are believed to have formed in the manner schematically outlined in figure 3.19, where the early stages of granite emplacement are relatively forceful, resulting in ductile shearing of the pelitic country rocks. This early deformation could be accommodated within the quartzites by shearing along the zone of dislocation, located just to the east of the quartzite - mobilised country rock boundary (see Fig. 3.19). Pitcher (1953)

and Rickard (1963) reported the occurrence of microgranite sheets in this area which they believed to be related to the Main Donegal Granite, but the present author can see no reason why these are not related to the Thorr Granite, since a similar sheet of relatively fine-grained granite that intrudes the mobilised pelites was found to be Thorr-type Granite, and not Rosses Granite as proposed by Pitcher (1953) and Pitcher & Berger (1972) (see Ch. 4). The rather chaotic nature of the folds in the field may reflect a general lack of coherence due to mobilisation, together with varying degrees of incompetence between pelitic and semipelitic horizons. The observed folding of the mobilised fabrics within this section, folding of andalusite, with the subsequent mimetic growth of sillimanite and folding of porphyroblast of sillimanite, lend support to the new hypothesis that the main phase folds in the migmatites are locally developed later structures in relation to forceful intrusion synchronous with thermal metamorphic dehydration and mobilisation of the country rocks (see Ch. 2). The main phase folds in the migmatites are, therefore localised syn-emplacement structures and they cannot be correlated with regional deformation events, thereby illustrating the risks of correlating deformation structures across large areas, especially in zones of active magmatism. The kinematics and timing of folding relative to final granite emplacement are discussed below.

It is unclear whether the early folds seen within the mobilised country rocks are F_2 folds or F_3 structures that have been tightened during folding in association with the early stages of granite intrusion. They may even be a combination of both F_2 and F_3 structures.

3.3.2 *Syn-emplacement deformation structures*

3.3.2.a *The mobilised country rocks*

The earliest syn-emplacement deformation structures seem to be the curvilinear and curvilinear main phase folds seen within the mobilised country rocks. The whale-back geometry and antiform to the NE vergence (Fig. 3.16) imply an early SW side up sense of shear (Fig. 3.19), which can be attributed to marginal shearing during early upward emplacement of the granite magma. This early phase seems to be overprinted by later sinistral shear because sinistral shear bands overprint the main phase of folds in the migmatites. These are filled with fine-grained granite similar in composition to the marginal facies of the

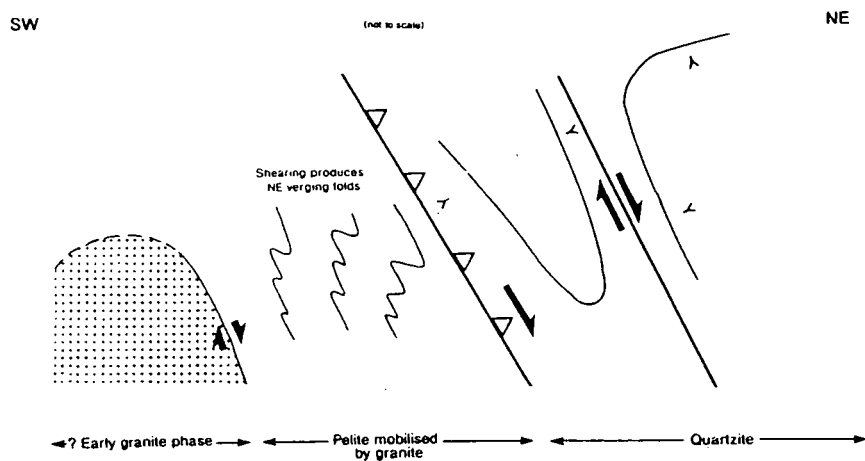
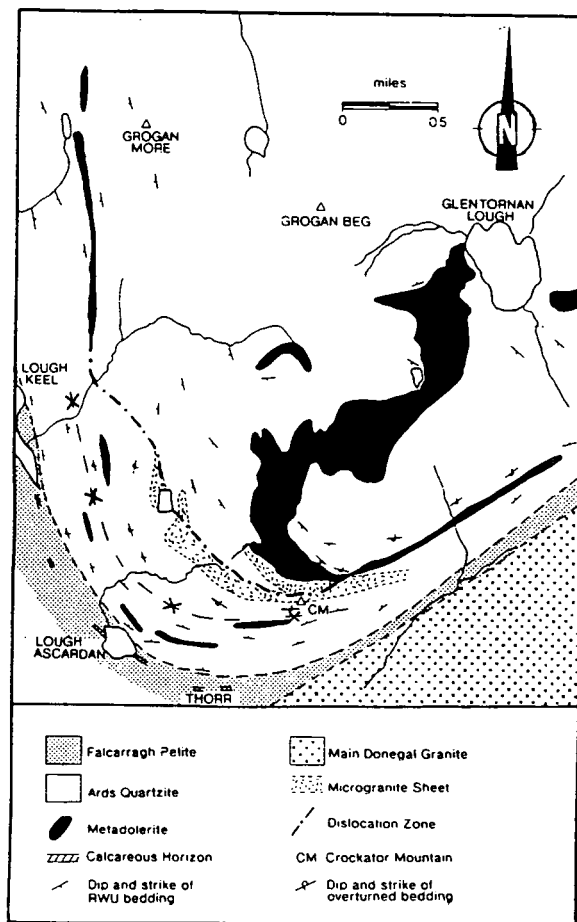


Figure 3.19 The strike swing in the Thorr area. (a) Map illustrating the major structural elements in the Thorr area; (b) sketch diagram to explain the origin of the main phase folds in the mobilised country rocks in relation to early forceful emplacement of the Thorr Granite. West-side-up shearing is accommodated by folding in the pelitic lithologies and by movement along pre-existing dislocations in quartzite.

pluton (see Ch. 2). The sinistral shear bands occur prolifically throughout the entire zone of mobilised pelites and semipelites, increasing in occurrence towards the main contact with the granite (see Maps 1 & 2). They have geometries consistent with them being shear fractures and the fact that they are filled with granite may mean that they are hydraulic shear fractures. According to Etheridge (1983), fractures like this indicate that the differential stress is $\leq 8T$, where T is equal to the tensile strength of the material. If this is applied within the mobilised country rocks at Thorr, the apparent differential stress, and hence the apparent ultimate strength of the mobilised pelites and semipelites is $\leq 8T$. This appears to differ from the granite veins seen in unmobilised country rocks (see below).

Other evidence for sinistral shear is seen at the quartzite - mobilised hornfels boundary. Here, early down-dip stretching lineations, that are seen clearly on the quartzite bedding surfaces, are rotated with an anticlockwise direction, to become strike-slip lineations. Thus, sinistral movements overprint dip-slip structures. Strike-slip lineations are also preferentially developed within thin pelitic horizons that are interbanded with the quartzites along the tectonic contact (see section 3.1.7.b). In addition, thin granite veins are augened to give top to the left, that is sinistral, sense of shear.

In summary, deformation structures related to granite emplacement are extensively developed within the mobilised hornfels next to the contact with the Thorr Granite in the southeast of the pluton. The earliest of these are non-cylindrical folds apparently formed by SW-side-up shearing along the early magma body margin. These dip-slip, forceful emplacement structures are subsequently overprinted by sinistral shear fractures that are filled with granite similar to the marginal facies of the intrusion. All these structures post date the regional F_2 and F_3 folds.

3.3.2.b *Emplacement related deformation structures elsewhere in the country rocks*

Deformation structures related to granite emplacement in the country rocks elsewhere around the pluton are mainly granite veins with sharp margins that intrude the country rocks by utilising pre-existing anisotropies such as bedding planes, foliation planes, cleavage planes and fold axial planes. Granite veins are particularly well developed within the quartzite country rocks, for example at Binanea Strand and Tory Island, where they are often seen to intrude along bedding planes. At these localities opening

directions for the veins (determined by matching marker horizons across fractures) consistently indicate that they are tensile fractures, with no shearing parallel to the vein walls. This observation is supported by the fact that the crystals within the veins form and grow at right angles to the veins walls. According to Etheridge (1983), veins of this type indicate that the differential stress ($\sigma_1 - \sigma_3$) associated with deformation is $\leq 4T$. There are two end-member explanations for the apparent difference in fracturing behaviour:

- the rocks in the southeast of the pluton (i.e. the mobilised hornfelses) represent a deeper level of the intrusion, where the external differential stresses are higher, (i.e. $8T$, not $4T$); or
- the difference in differential stress is only an *apparent* difference since the mobilised pelites and unmobilised country rocks, especially quartzite, may have very different ultimate strengths, reflecting the marked difference in their composition (i.e. a fixed level of $(\sigma_1 - \sigma_3)$ is $\geq 8T$ in the migmatites and $\leq 4T$ in the other regions).

Given the obvious difference in composition and response of the quartzite and pelite lithologies to thermal metamorphism, the latter option seems more likely. Nevertheless, the possibility that the southeastern contact lies at greater crustal depths cannot be ignored since evidence of shear fractures are seen in the rafts close to this area, in the form of isolated sinistral shear band granite veins (e.g. Ardveen and Lough Agher).

Conclusions

- (1) The country rocks surrounding the Thorr pluton experienced significant deformation during the Caledonian orogeny.
- (2) The type of structures formed in response to deformation vary according to lithology, since relatively incompetent pelitic lithologies can undergo ductile deformation more easily than relatively competent quartzites.
- (3) In each of the subareas defined around the pluton (see introduction to this chapter) deformation structures can be subdivided into those that pre-date granite intrusion and those that are synchronous with it.
- (4) In most subareas two sets of folds pre-date granite intrusion. The earliest of these are correlated with regional F_2 folds and the main phase folds in each subarea (with the exception of the main phase folds in the mobilised hornfelses) are correlated with regional F_3 folds.
- (5) The main phase folds in the mobilised Falcarragh Pelites, in the southeast of the pluton have different vergence directions and slightly more chaotic geometries than the main phase folds in the rafts and at Maghery. They fold metamorphic index minerals such as andalusite, fibrolite and sillimanite; these minerals randomly overprint fabrics associated with the main phase folds elsewhere around the pluton. It is, therefore, proposed that the main phase folds within the mobilised hornfelses post-date the regional F_3 folds, and are related to early forceful intrusion of the Thorr Granite in this area in the southeast of the pluton. This is the only evidence of forceful intrusion of the Thorr granite.
- (6) The main phase folds in the mobilised hornfelses are overprinted by sinistral shear bands and these are believed to have formed during the main phase of granite emplacement. Rough estimates of the amount of sinistral motion across the mobilised zone are a minimum of 5m and a maximum of 250m.
- (7) There is also evidence of sinistral shearing synchronous with granite emplacement along the tectonic contact between mobilised Falcarragh Pelites and unmobilised Ards Quartzite, to the east of the main granite contact in the SE of the pluton.

(8) Emplacement of granite veins elsewhere around the pluton was apparently passive and involved tensile fracturing utilising pre-existing anisotropies in the country rocks, including bedding planes, foliation planes, cleavage planes and fold axial planes.

(9) All these features suggest that hydraulic fracturing associated with high fluid pressures played a significant role in bringing about the emplacement of the Thorri Pluton.

(10) The difference in mode of fracturing during granite emplacement probably reflects the strong rheological differences between the highly mobilised migmatites in the SE of the pluton and the more coherent country rock lithologies elsewhere. However, these results might also reflect differences in original depth, with the SE contacts of the pluton exposing originally greater depths that were, therefore, subjected to larger differential stresses.

Chapter 4

THE IGNEOUS FACIES OF THE THORR GRANITE, THEIR DEFORMATION FABRICS AND TEXTURES

Introduction

The detailed petrography of the igneous facies has been described by several previous workers (e.g. Pitcher 1951, 1953a; Whitten 1957; Pitcher & Berger 1972; Oglethorpe 1987; see summary in Ch. 2) and will not be re-examined in rigorous detail here. However, representative samples have been selected from designated subareas to give a brief guide to the petrographical variation within the Thorr Pluton. Samples will be described from the following subareas:

Tory Island

Binanea Strand

Tievelehid (not shown in Fig. 4.1, see Map 2)

Inishbofin Island

Curran's Point

NW Coast (N of Maghera Strand)

W Coast (mainland opposite and including Gola Island)

SW Coast (Annagry - Burtonport)

SE margin (Crolly - Meencorwick - Lough Agher - Ardveen - Lough Anure)

Maghera - Toberkeen

Aranmore Island

Lettermacaward (not shown in Fig. 4.1, see Map 2)

Cleengort Hill (not shown in Fig. 4.1, see Map 2).

In addition, unusual rock types, such as the appinites of the north coast and the orbicular granite of Mullaghderg will be described briefly.

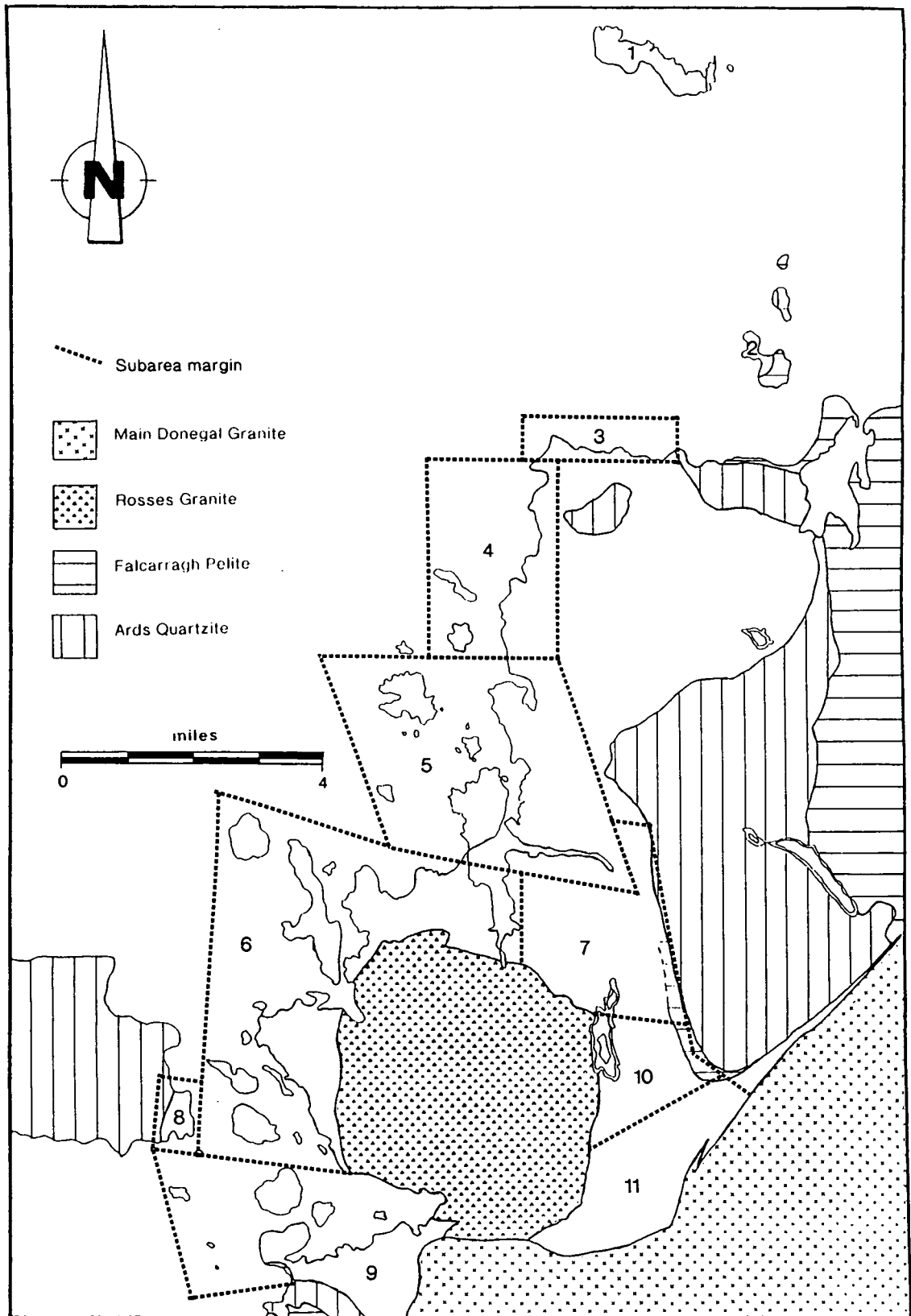


Figure 4.1 General map of NW Donegal illustrating the position of most of the subareas mentioned in this chapter (for remaining subareas see Map 2).

4.1 *The field relationships and petrography of the granites and granodiorites*

The petrography of the whole of the Thorr Pluton was described by Oglethorpe (1987); prior to this, petrographical studies of the pluton concentrated on particular subareas, with Pitcher (1951, 1953a) describing the petrography of the Thorr area, whilst Whitten (1957a, b) described the petrography of the more northerly regions, from Gweedore to Tory Island.

A typical specimen of the Thorr granite is illustrated in Plate 4.1. It is coarse grained and contains large pink alkali feldspar phenocrysts, plagioclase, biotite, hornblende and sphene. The margins of the pluton tend to be more mafic than the central parts and alkali feldspar megacrysts are less common (Pitcher 1951, 1953a; Whitten 1957b; Oglethorpe 1987; Ch.2.6). The most basic examples of Thorr Granite occur as xenoliths within and along the northern margin of the Main Donegal Granite. The central part of the Thorr Pluton, as seen around the towns of Bunbeg and Derrybeg, have lower colour indices (≤ 10) due to a lack of mafic minerals, in particular hornblende (see Ch. 2.6). The granite here is rich in alkali feldspar and quartz. Whitten (1957b) called this part of the Thorr Pluton the 'Gola Granite', postulating that it was of late metasomatic origin.

Detailed field studies of the Thorr Granite have shown that it is very heterogeneous. There are impersistent internal contacts between granites (*sensu lato*) of very different colour index and of differing grain size. In addition, there are also ellipsoidal- or discoidal-shaped basic 'patches' which tend to be found in swarms. Some of these may be metasedimentary xenoliths, but others (for example the many examples seen at Rinardalliff Point on the Bloody Foreland) are certainly cognate inclusions of more basic magma types. These maybe related to the appinitic and dioritic magmas presently seen in faulted contact with the granite along the north coast (see section 4.2 & Map 2).

The following subsections give brief petrographical descriptions of thin sections of the Thorr Granite in subareas, roughly from N to S throughout the pluton. Plates of representative granite facies are included.



Plate 4.1 Typical appearance of an outcrop of Thorr Granite.

4.1.1 *Tory Island*

The Thorr Granite outcrops in the west of this island (Map 2). The contact between the granite and the quartzitic country rocks is poorly exposed inland, but can be seen in the steep cliffs of the north coast, at Arderrill, and in the low tidal reefs at Torranaman, trending roughly NW-SE. Where exposed, the contact is generally sharp, but often transgresses the bedding surfaces. In the low tidal reefs at Torranaman, south of the main contact, large randomly orientated blocks of quartzite are enveloped in the coarse grained granite. The quartzite country rocks east of the main contact often contain metre-scale granite veins of similar composition to the typical granite seen on the island (see Ch. 3) and thinner veins of microgranite in interlocking networks.

The mineralogy of the granite on Tory Island is as follows; plagioclase feldspar, alkali feldspar (microcline- microperthite), quartz, biotite, hornblende and sphene. Microcline microperthite often completely encloses small, aligned plagioclase crystals. In addition to this fairly typical Thorr Granite, NW-SE trending, impersistent sheets of both fine grained, mafic-rich material and alkali feldspar megacrystic granite outcrop on the north coast of the island (NC 856468, Map 2), almost due north of East Town, the main town on the island. These sheets cannot be traced for more than 50 metres, partly because of the poor inland exposure, and as equivalent units do not outcrop on the south coast, they are therefore believed to be localised, impersistent sheets.

4.1.2 *Binanea Strand*

The granite contact is poorly exposed here, but specimens of granite were collected from as close to the transition from quartzite country rocks as possible. Whitten (1957a) stated that the contact here is subvertical, since it is straight and descends from an elevation of 350 feet to sea level. Close to the contact (NC 893312, Map 2), a small (c. 5-10m) outcrop of intrusion breccia is well-exposed. This contains angular fragments of quartzite (up to 25cm in diameter) and occasional semipelite and micaceous quartzite, set in a leucocratic granitic groundmass. It is similar to a smaller intrusion breccia seen in the well-exposed quarry section south of Lough Keel in the Thorr area. The present author agrees with the conclusion of Whitten (1957a) that such breccias are associated with granite intrusion. At Binanea Strand

three sets of tensile granite veins are developed. The main set lies subparallel to the bedding in the quartzite, and the others dip more steeply than bedding and apparently form a conjugate set (see Ch. 3).

The main granite is mostly medium grained and is exposed in patchily distributed outcrops at this NE end of the pluton. Close inspection with a hand lens often reveals irregular intergrowths of quartz and feldspar, for example at NC 895309, just south of the main quartzite - granite contact above Binanea strand. This is well-illustrated in thin sections of specimens taken from this area (see Plate 4.2). Textures such as this often develop at higher crustal levels (c.f. the Ben Stumanadh Syenite, Ch. 6), due to relatively rapid decrease in vapour pressure, which results in eutectic crystallisation of quartz and feldspar (Mason 1985). The granite in this region is relatively rich in quartz and feldspar, and contains less biotite and hornblende. Quartz can sometimes account for 40 - 50 % of the total felsic minerals. The texture is more equigranular compared to other parts of the pluton, containing fewer phenocrysts and a weaker deformation fabric (see section 4.3).

Evidence for metre- to centimetre-scale interdigitation of quartzite and granite can be observed inland and W of Binanea Strand (see Map 2, NC 887316). Steeply dipping, NE-trending granite sheets, up to about a metre wide, cross-cut the ESE-striking and shallowly dipping quartzite bedding. These granite veins are rather fine grained, but contain phenocrysts of feldspar (both plagioclase and alkali feldspar) and biotite, that are preferentially aligned roughly parallel to the vein walls. Plate 4.3 is a photomicrograph of a specimen taken from one of these granite veins. The groundmass is almost exclusively fine grained quartz that exhibits an undeformed igneous texture. Phenocrysts of feldspar and biotite exhibit a well-developed shape preferred orientation within the undeformed groundmass that is typical of magmatic state deformation fabrics (see sections 1.3.1 & 4.3). Cooling of the magma must have been relatively rapid in order to produce a texture like this and the present author interprets this as further evidence that granite emplacement took place at relatively shallow crustal levels in this NE part of the Thorr Pluton.

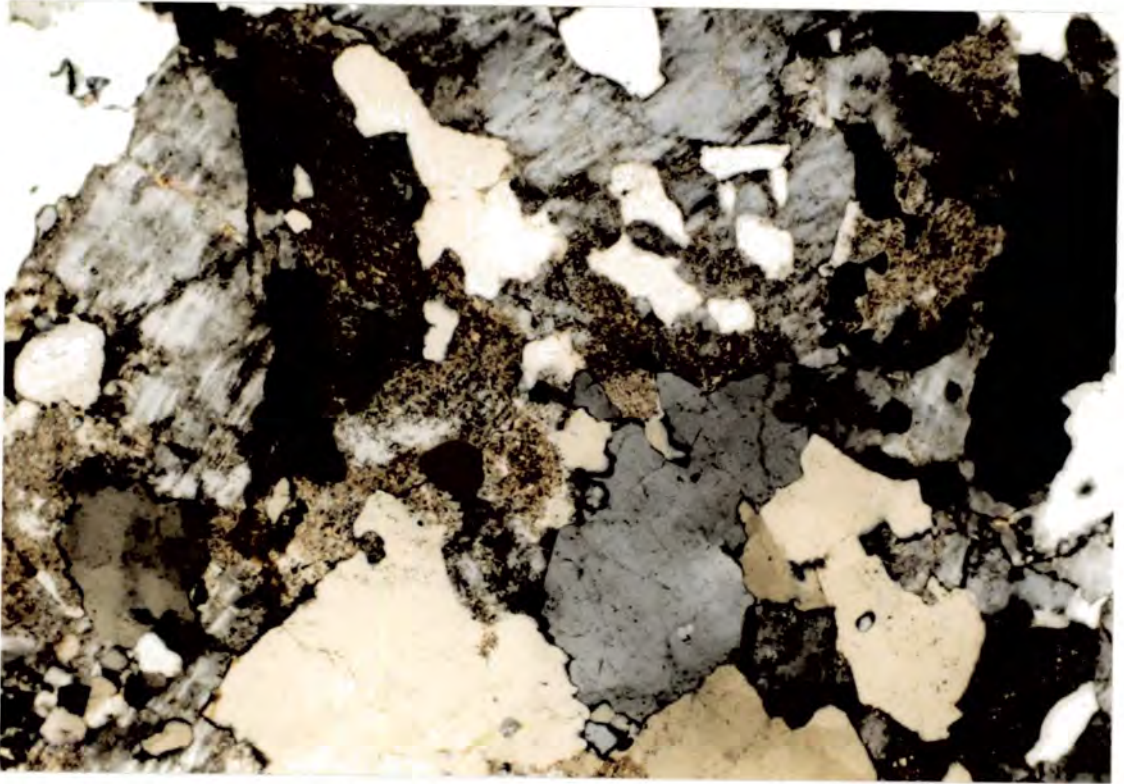


Plate 4.2 Photomicrograph of specimen of Thorr Granite collected close to the granite - quartzite contact at Binanea Strand (NC 895 309). Photograph illustrates the intergrowth of quartz and feldspar. (Field of view 4.4mm).

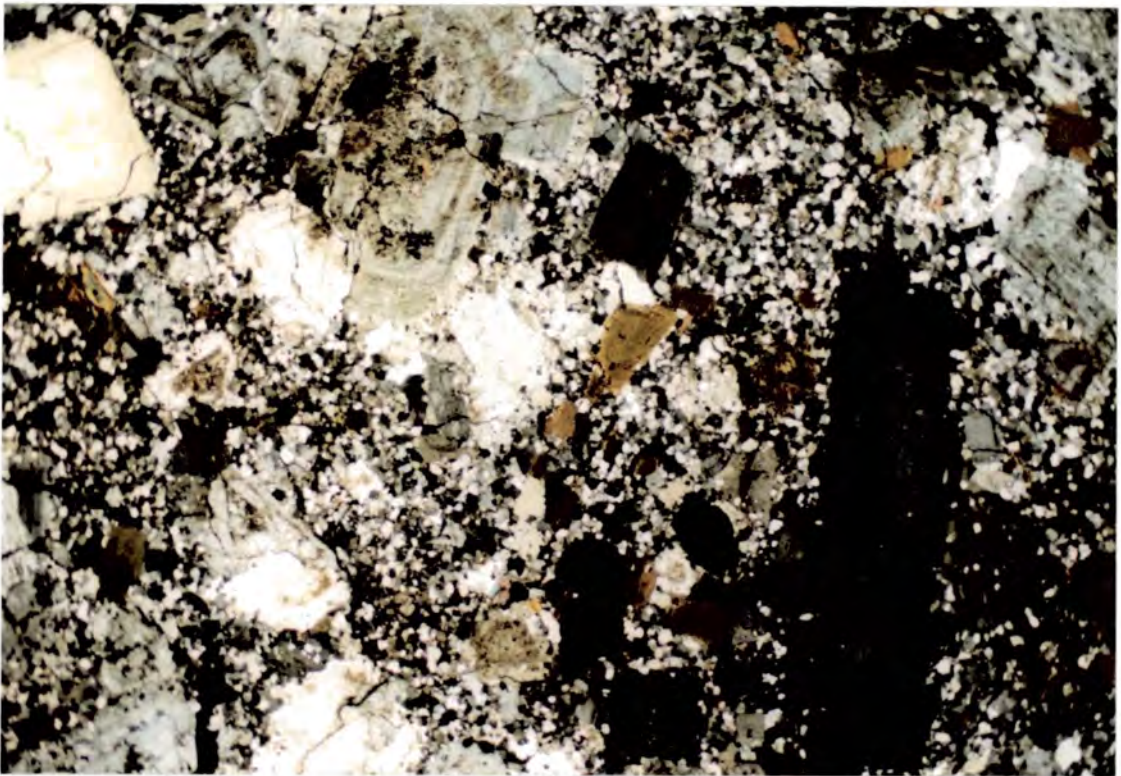


Plate 4.3 Photomicrograph of specimen of granite vein collected at NC 887 316. The photograph illustrates the porphyritic texture typical of granite veins in this area. Phenocrysts of feldspar and mafic minerals are set in a fine-grained, equigranular quartz matrix. (Field of view 18mm).

4.1.3 *Tievelehid*

Lithologies are generally very poorly exposed in this part of the pluton, due to a thick cover of both boulder clay and peat. Granite was found to outcrop in a few stream sections (see Map 2). In the field it has a red, rusty colour and crumbles easily when struck with a hammer, so the specimens collected are rather weathered and altered. In thin section, it contains biotite and hornblende, both of which have been altered to chlorite, as the main mafic phases, along with accessory sphene. Felsic minerals are microcline microperthite, plagioclase and quartz. Plagioclase exhibits very well developed oscillatory zoning. Microcline microperthite encloses small, aligned plagioclase minerals and is itself aligned in an equant groundmass of quartz. All of these textural elements combine to form a well developed magmatic state deformation fabric (see Ch. 1).

4.1.4 *Inishbofin Island*

The granite on this island exhibits similar features to that on the mainland near the southeast of the pluton, but on a slightly smaller scale. Generally, the colour index increases towards the contact with semipelitic country rocks. Thin sections of the granite close to the contact (Plate 4.9) reveal that the locally developed 'contact facies' contains red biotite typical of Thorr contact facies on the mainland (c.f. Ch. 2.6). In addition, sillimanite, muscovite and ordinary biotite intergrown with hornblende, form common mafic 'clots' close to the contact with the country rocks. Furthermore, the relative proportion of quartz is greater towards the contact, often at the expense of feldspar. Early magmatic state deformation fabrics, defined by feldspar phenocrysts and euhedral biotite and hornblende, are overprinted by solid state fabrics of variable intensity. The development of the solid state fabrics tends to be most intense close to, or within the mafic clots, and quartz and mafic minerals within and around these clots sometimes develop annealed textures. Away from the contact, the granite carries less quartz or biotite and hornblende, and is more typical of the normal facies of the Thorr Granite (Oglethorpe 1987; see Ch. 2.6) that is exposed along the NW coast of the mainland.

4.1.5 Curran's Point

The granite exposed here is not typical of the normal facies (Oglethorpe 1987) Thorr Granite. There are two main reasons for this; firstly, there is a concentration of basic magmas from Curran's Point, eastwards along the north coast, to Magheraroarty (see Map 2). Secondly, the granite is seen in contact with pelitic and semipelitic lithologies, resulting in the development of narrow zones ($\leq 10\text{m}$ wide) of contact facies Thorr Granite (*sensu* Oglethorpe 1987) that are more abundant in mafic minerals than the normal facies granite. The relative abundance of mafic magmas seems to be of particular importance and Plate 4.4 illustrates just how abundant biotite and hornblende is within a fairly typical specimen from Curran's Point that has been contaminated by mafic magmas. The nature of the mafic rocks that outcrop along the north coast will be described in more detail in section 4.2 of this chapter.

4.1.6 NW Coast - north of Maghera Strand

Granite exposed along this coast is typically coarse grained and has abundant feldspar phenocrysts, that define a well developed shape preferred orientation typical of a magmatic state deformation fabric. In thin section, this granite is typically mafic-poor, containing very little biotite and virtually no hornblende. Previous investigators (Whitten, 1957b; Oglethorpe 1987) reported the occurrence of centimetre- to metre-scale, impersistent sheets of both fine to medium grained granite, granodiorite and diorite along the NW coast. Plate 4.5 is an example of such a sheet of relatively fine grained granodiorite exposed along this coastline, near Bunaniver Port (see Map 2). These sheets generally strike parallel to the preferred orientation of feldspar phenocrysts, which is roughly NW-SE in this area, but locally they are openly warped, or folded on a metre-scale.

4.1.7 W Coast - Bunbeg to Maghera Strand, including Gola Island

Plate 4.6 is a photomicrograph from a representative sample of the granite in this area. It is a relatively felsic facies of the Thorr Granite, with only small proportions of mafic minerals. This specimen was collected on Gola Island where the granite is devoid of hornblende (see Fig. 2.8 for the extent of the hornblende-free granite). It is representative of what Whitten (1957b) called the Gola Granite, and

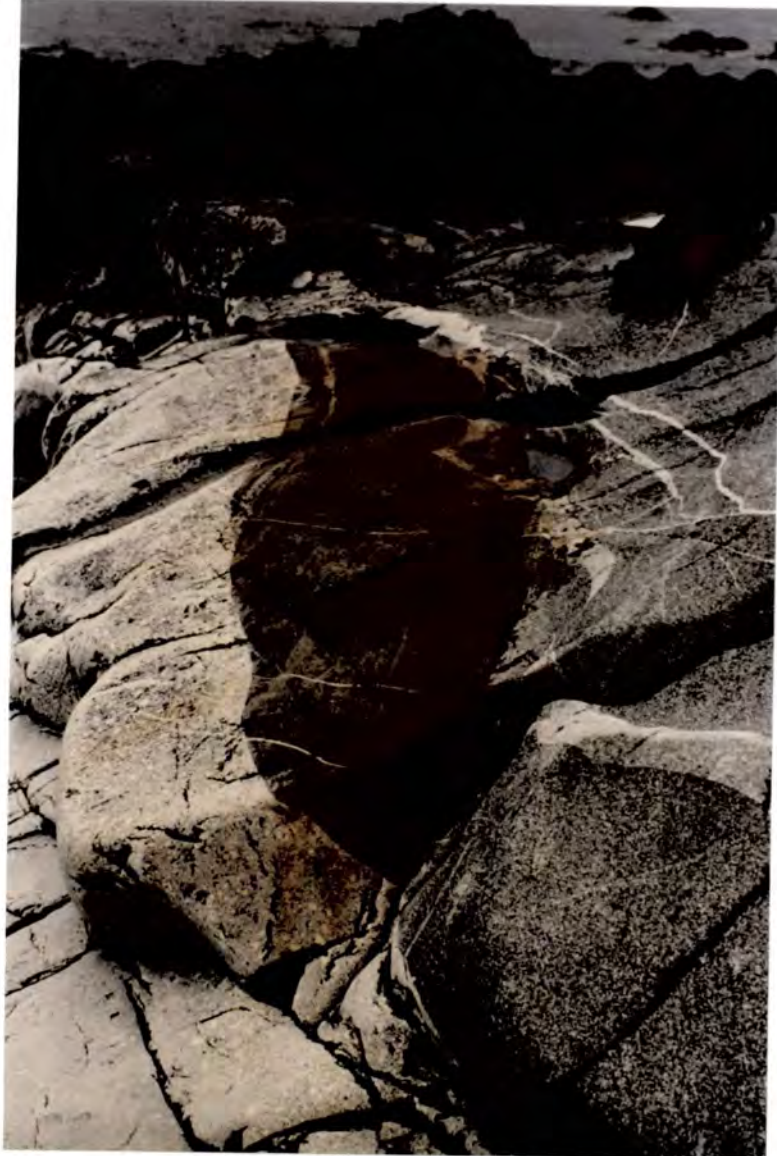


Plate 4.4 Outcrops of relatively melanocratic granodiorite/diorite at Curran's Point. Close inspection shows just how abundant mafic minerals are in this lithology.



Plate 4.5 Example of a thin co-magmatic dioritic dyke within Thorr Granite.

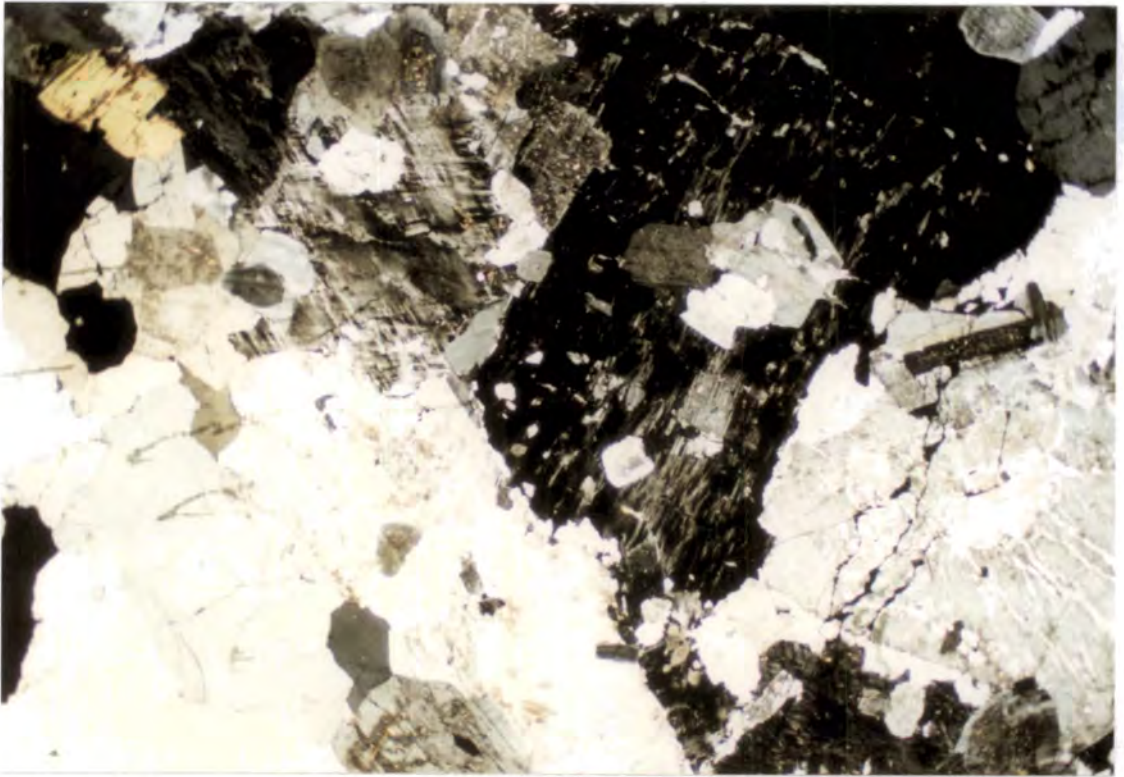


Plate 4.6 Photomicrograph of a representative sample of 'Gola facies' Thorr Granite. Note the relative abundance of felsic minerals and the lack of mafic minerals, especially hornblende. (Field of view 18mm).

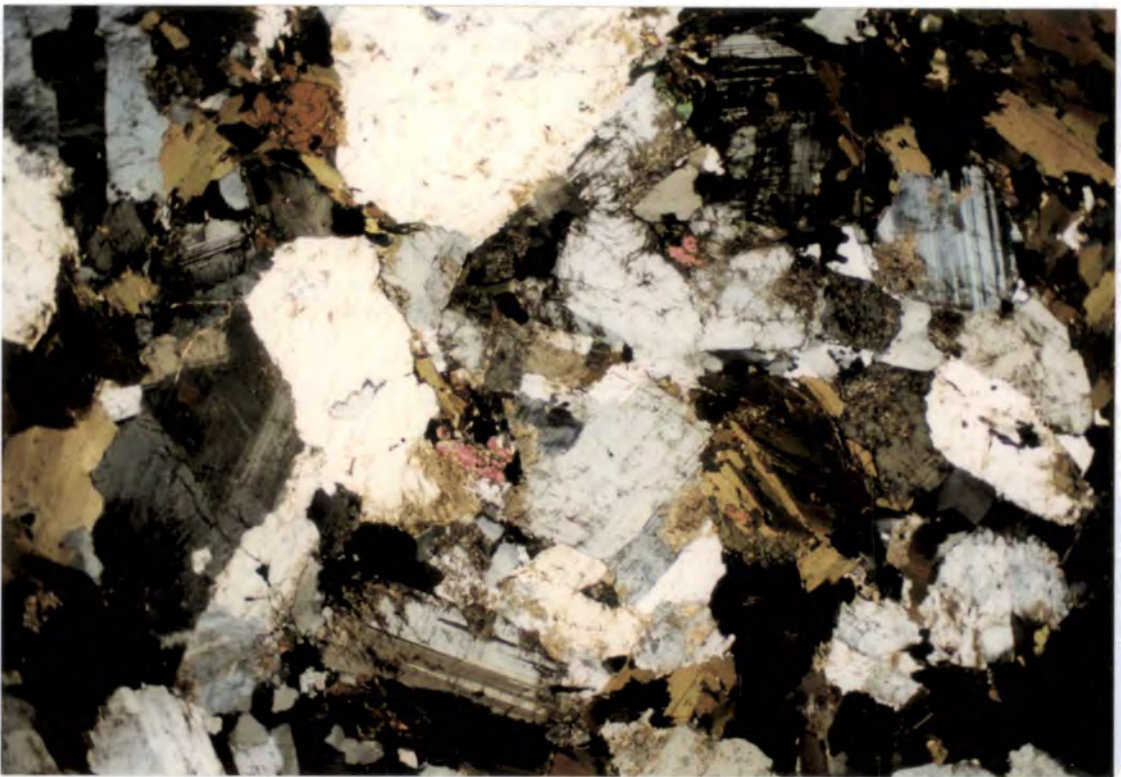


Plate 4.7 Photomicrograph of relatively melanocratic 'normal facies' Thorr Granite. Note also the euhedral, aligned phenocrysts; these define a well developed magmatic state deformation fabric. (Field of view 18mm).

Oglethorpe (1987) called the Hornblende-free normal facies (see Ch. 2.6). Its mineralogy is dominated by feldspar phenocrysts, both plagioclase and microcline microperthite, that are aligned to form a shape preferred orientation in a groundmass of relatively undeformed quartz, and are, therefore, typical magmatic state deformation fabrics. Quartz grains are usually very large and show only minimal amounts of plastic deformation in the form of undulose extinction and minor sub-grain development.

4.1.8 SW Coast - Annagry to Burtonport

The mineralogy of these rocks varies from a typical assemblage representative of the normal facies Thorr Granite (Oglethorpe 1987) through to a more felsic facies, with very few mafic minerals. However, the Gola facies granites of this subarea are never completely hornblende-free, although some may contain less than 1% modal quartz (Oglethorpe 1987; also see Fig. 2.8). Plate 4.7 is a photomicrograph of a relatively mafic facies of the granite, with greater than 1% modal hornblende. Hornblende is often intimately intergrown with biotite. There are also well-developed phenocrysts of both plagioclase and alkali feldspar that have a strong shape preferred orientation. Occasionally, myrmekite is developed at the extreme edges of the feldspar phenocrysts, especially where alkali feldspar is in contact with plagioclase. Quartz is typically interstitial to all these phenocryst phases. It forms large anhedral grains, many of which show undulose extinction and subgraining, and occasionally a little grain size reduction due to dynamic recrystallisation of large grains into smaller grains. The extent of this plastic deformation of quartz is very variable, but is never pervasive (see section 4.3). The granite here is very altered, with hornblende and biotite altered to chlorite and feldspar heavily sericitised, this is typical of the Thorr Granite over the whole extent of the pluton.

4.1.9 The SE margin - Crolly, Meencorwick, Lough Agher, Ardveen, Lough Anure.

The granite in this area shows considerable mineralogical variation, apparently due to interaction with the pelitic and semipelitic country rock lithologies that are found next to the main granite contact and as xenolithic rafts within the granite. Many previous workers have indicated that distinctly different granite facies exist in this area. Pitcher (1951, 1953), Pitcher & Berger (1972) and Scott (1974) proposed

that the variation in composition at Thorr and Ardveen could be attributed to contamination of the magma due to assimilation of pelitic and semipelitic country rocks. Oglethorpe (1987) suggested that wholesale assimilation of metasediments was unlikely on chemical grounds, proposing instead that preferential exchange of elements between the country rocks and the granite was a more likely process (see Ch. 2.6). Although the Thorr Granite in this area contains many mafic inclusions and schlieren that appear to be remnants of digested country rock, trace element studies (Sm-Nd & Rb-Sr) of the Thorr Pluton and other Donegal Granites (Dempsey *et al.* 1990) indicate that the Thorr Granite has a relatively primitive chemistry, rather typical of a melt derived from the mantle, with little or no crustal input. This data corroborates that of Oglethorpe (1987) and suggests that wholesale contamination of the Thorr magma by partial melts of metasedimentary country rocks did not occur to any great extent.

Plates 4.8, 9 & 10 illustrate the salient features of the mineralogy of the normal, transitional and contact facies of the Thorr Granite close to the main contact and around the metasedimentary rafts. Plate 4.8 is a typical coarse grained normal facies granite from this area. Its colour index is somewhat higher than the normal facies seen further north and west, since it contains more biotite and hornblende. Biotite is typically green-brown and hornblende bluish-green. They tend to occur together as aggregates and are often intergrown with one another. Both biotite and hornblende are subhedral to euhedral. This specimen also contains phenocrysts of plagioclase feldspar and alkali feldspar (microcline microperthite), although the latter is not as abundant as further north in the pluton. Quartz is typically interstitial and is relatively equant. It exhibits only limited evidence of having been plastically deformed, having undulose extinction and with the development of subgrains and, in extreme cases, recrystallisation to smaller new grains. Even this 'uncontaminated' granite phase is not completely homogeneous in the field. At Ardveen, impersistent N-NNW trending, internal contacts between coarse grained and medium grained variants can be seen. Scott (1974) interpreted this as evidence that the Thorr magma was not intruded as a single pulse, but perhaps as a series of surges of magma.

Plate 4.9 illustrates a typical transitional facies (see Ch. 2.6) of the Thorr granite, which is found up to 10-15m away from rafts of pelitic and semipelitic country rock. The transitional facies is usually medium to coarse grained and mineralogically it is characterised by higher quartz contents and limited

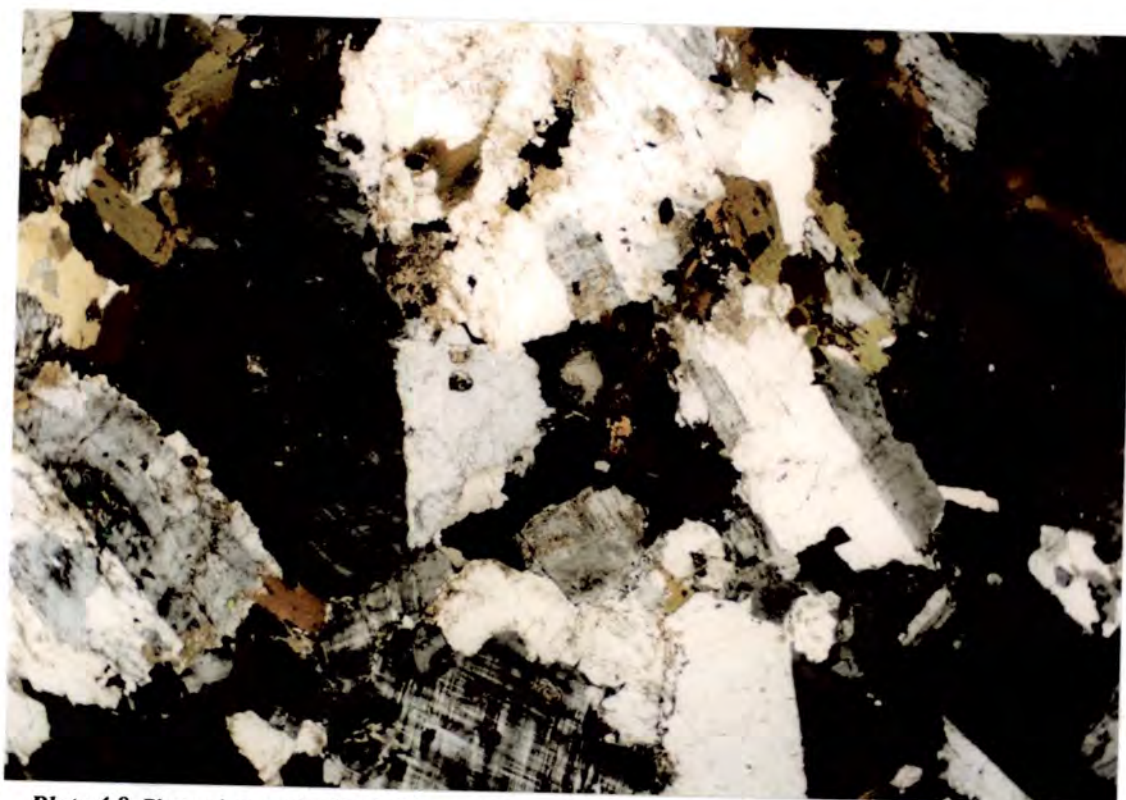


Plate 4.8 Photomicrograph of typical 'normal facies' Thorr Granite from the SE of the pluton. Note that the majority of biotite in this specimen is typically green-brown in colour. (Field of view 18mm).



Plate 4.9 Photomicrograph of typical 'transitional facies' Thorr Granite from around a country rock raft in the SE of the pluton. Note the occurrence of two varieties of biotite, one with the usual green-brown colour and the other with a foxy red-brown colour. (Field of view 18mm).

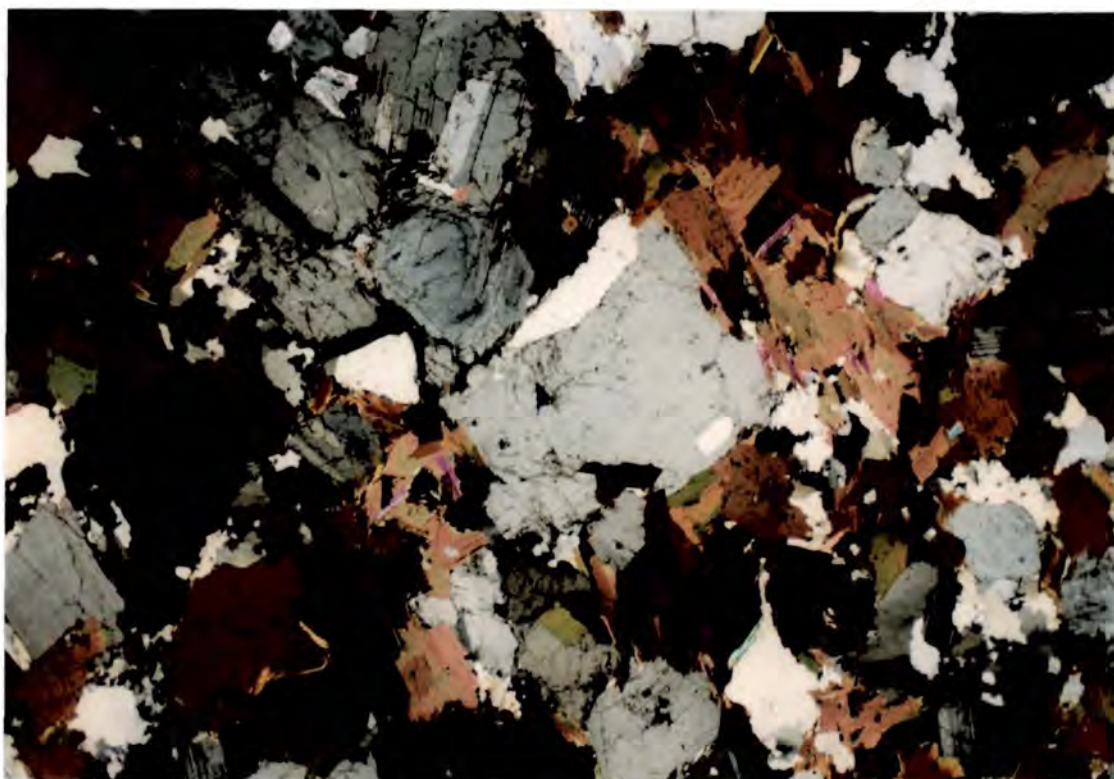


Plate 4.10 Photomicrograph of a typical 'contact facies' Thorr Granite from adjacent to a the main contact in the SE of the pluton. Note the pervasive occurrence of the foxy red-brown variety of biotite, which is more common than ordinary green-brown biotite. (Field of view 18mm).



Plate 4.11 Appearance of a typical outcrop of contact facies granite. Note the finer-grained and more equigranular appearance, the abundance of mafic minerals and the 'knotty' quartz grains that tend to have a bluish tinge.

development of strongly coloured, foxy red biotite (see Ch.2.6), in addition to normal green-brown biotite. The transitional facies grades smoothly into the medium grained contact facies rocks found immediately next to country rock rafts and adjacent to the main granite contact. Plate 4.10 is a photomicrograph of a specimen of the contact facies. Plate 4.11 illustrates its appearance in the field. Both of these photographs show that the contact facies is relatively mafic rich. Foxy red biotite is abundant (clearly illustrated in plate 4.10) and contains euhedral inclusions of apatite with dusky cores. In addition, the contact facies has abundant, knotty, interstitial quartz, this is easily observed in the field. Oglethorpe (1987) attributed the relative abundance of quartz within the contact facies to enrichment of the magma in silica derived from the dehydration of the pelitic and semipelitic metasediments and carried in hydrothermal fluids. The textures of the granites in this area will be described in section 4.3 of this chapter, since they appear to have involved more solid state deformation than elsewhere in the pluton.

4.1.10 *Maghery - Toberkeen*

The granite here is seen in contact with Ards Pelite (Meneilly 1982) and Ards Quartzite (Pitcher & Berger 1972; Meneilly 1982). In addition, numerous xenolithic rafts of country rock, mainly pelite and semipelite, are included within the granite in this subarea south of Dungloe. The granite around these rafts and at the main contact exposed north of Maghery Bay, where it is seen in contact with Ards Pelite, is similar to the contact and transitional facies around Meencorwick, described in the preceding subsection. The contact facies at Maghery, like that at Meencorwick, has a relatively high colour index and carries strongly coloured (foxy red) biotite in addition to green-brown biotite and blue-green hornblende. In addition, the quartz content is relatively high. Two types of feldspar are seen in the contact facies at Maghery, but microcline is not as abundant as in the normal facies granite. Away from the pelitic country rocks and rafts, the granite is more felsic and only contains green-brown coloured biotite, relatively less quartz and the proportion of alkali feldspar is higher than in the contact facies. No contact facies is developed where the Thorr Granite is seen in contact with Ards Quartzite (e.g. NC 742077). The granite inland of Maghery exhibits a variety of deformation textures which will be discussed in sections 4.3 & 4.4.

4.1.11 Aranmore Island

The granite exposed on this island is very similar to the coarse grained normal facies granite exposed in the SW of the pluton on the mainland and the 'uncontaminated' granite at Maghery. It is a feldspar-dominated rock, containing phenocrysts of plagioclase, which tend to be small and equant and microcline microperthite, which tends to occur as large euhedral to subhedral grains in thin section. Green-brown biotite and blue-green hornblende are seen as small, euhedral to subhedral grains and often form mafic 'clots' or aggregates. Sphene is a common accessory phase that is often associated with the mafic aggregates. It is always euhedral and commonly twinned. Quartz is interstitial to all of these euhedral phases and has the form of large grains with common undulose extinction and often subgrains.

4.1.12 Lettermacaward

Pitcher (1951, 1953) and Pitcher & Berger (1972) described the granite in this area as part of the 'Older Granodiorite', which includes the granite facies around Thorr and generally wherever it contains numerous pelitic and semipelitic country rock rafts. In the field it is rather mafic-rich and it contains large, knotty grains of quartz and is very similar in appearance to the contact facies granite described from Meencorwick. All the characteristic contact facies minerals are seen; foxy red biotite, often containing inclusions of apatite; large interstitial quartz grains with serrated grain boundaries; plagioclase and microcline microperthite; green-brown biotite which is often intergrown with blue-green hornblende; muscovite and sillimanite which are spatially associated with the strongly coloured biotite. Quartz grains often enclose small, euhedral mafic minerals and carry large fluid inclusions.

4.1.13 Cleengort Hill

The Thorr Granite in this area just south of the Gweebarra River is rather felsic in composition, containing only a small amount of biotite. The grain size here is very variable, from very fine to very coarse grained. In the field, the variation in grain size can be attributed to a concentration of garnet and sillimanite bearing aplite and pegmatite veins in this area. In addition, the grain size has been reduced due to plastic deformation of most of the mineral phases in response to deformation associated with the

later Main Donegal Granite Shear Zone (MDGSZ). The texture of the granite in this area will be discussed in sections 4.3 & 4.4.

4.2 Texturally and compositionally unusual intrusions within the Thorr Pluton

4.2.1 The Orbicular Granite

A small body of Orbicular Granite, 4.5 - 6m wide, outcrops near the low tide mark in the district of Mullaghderg (NC 75452105, Map 2). It was studied and has been described in detail by Hatch (1888) and Cole (1916). The observations of the above investigators were summarised by Pitcher and Berger (1972).

This granite facies is composed of a cluster of orbicules that have a range of diameters up to a maximum of 15cm (Plate 4.12). Each orbicule consists of an inner, feldspathic core and an outer, more mafic mantle. The whole body is contained within a thin, sack-like sheath of feldspathic material, which is similar in composition to the cores of the orbicules. In addition to the main body, individual orbicules are found in the coarse grained granite host, up to 10m away from it. The orbicules are sometimes rather ellipsoidal in shape. In the field this can be seen to be due the fact that they impinge on one another. This indicates that there may have been mutual interference of orbicules at some stage during their growth.

In hand specimen, the feldspathic cores can be seen to be up to 7cm in diameter and are composed of a coarse grained association of oligoclase, alkali feldspar, quartz and accessory sphene. The peripheral mantle to the orbicules is often composed of single crystals of oligoclase radiating outwards from the core to the boundary. In addition, mafic minerals, such as biotite and magnetite, are often concentrated within the mantle, where they are arranged in concentric 'shells'.

Pitcher and Berger (1972) reported that occasionally the cores of orbicules consist of baked pelitic material; this feature was not observed by the present author. Cole (1916) postulated on the origin of these infrequently observed xenolithic cores, suggesting that fragments of country rock may have been



Plate 4.12 An outcrop of orbicular granite exposed at the high tide mark near Mullaghderg (NC 7545 2105).

incorporated in the magma, and once there, they played a significant role in nucleating the growth of feldspar in the radial form typical of the orbicules.

Pitcher and Berger (*loc. cit.*) suggest two alternative hypotheses for the production of such a body, both of which involved crystallisation from a magma that was virtually crystal free, but requiring a rather delicately balanced system. They envisaged a "bag of immiscible globules of viscous magma" that interfered with one another early in their growth history, but finally consolidated by individually undergoing radial crystallisation.

4.2.2 *The appinitic and dioritic rocks*

These are concentrated around the NW and W coasts, particularly along a section from Curran's Point to Meenalargan (Map 2). Curtis (1959) carried out detailed petrological studies of these basic rocks and Oglethorpe (1987) briefly analysed their geochemistry.

In the field and in hand specimen, there is an obvious increase in the proportion of hornblende, with a corresponding decrease in grain size towards the north of the pluton. At the same time, impersistent dioritic streaks, synplutonic dykes and cognate xenoliths become increasingly abundant.

The basic rocks generally occur either as large, amorphous bodies, or centimetre- to metre-scale synplutonic dykes, or centimetre-scale cognate xenoliths. The amorphous bodies are restricted to the section from Curran's Point to Meenalargan (Map 1), whilst dykes and enclaves are more widespread, occurring along the length of the NW and N coasts. The basic rocks range in composition from almost exclusively hornblende, pyroxene and biotite bearing rocks (Plate 4.13), through to coarse grained hornblende granodiorites (Plate 4.14). In addition, at several localities along this section of the north coast (NC 886336 & 867335), basic rocks outcrop as explosion breccias (Plate 4.15). These, like other intrusion breccias seen within the Thorr Pluton (e.g. Derryconor and Marley's Rocks) are believed to be the products of fluidization due to gas action (Whitten 1959a). Generally, however, the field relationships between diorites, appinite and granites indicate that they were emplaced synchronously. This is particularly well illustrated in metre-wide dykes, which often have very sharp contacts with the granitic host. In detail, however, many of the margins are crenulate (Plate 4.16), indicating that the contacts



Plate 4.13 Outcrop of a hornblende, pyroxene, biotite and feldspar bearing rock along the north coast of NW Donegal, east of Curran's Point.



Plate 4.14 Hornblende granodioritic facies exposed within the 'appinites' along the north coast of NW Donegal.



Plate 4.15 Example of an explosion breccia within the appinitic body exposed along the north coast of NW Donegal. The breccia consists of very angular clasts of medium-grained diorite set in a matrix of medium-grained granodiorite.



Plate 4.16 Contact between dioritic and granitic rocks at Curran's Point. Generally the contacts are sharp, however, in detail they are crenulate and typical of contacts between two immiscible liquids.

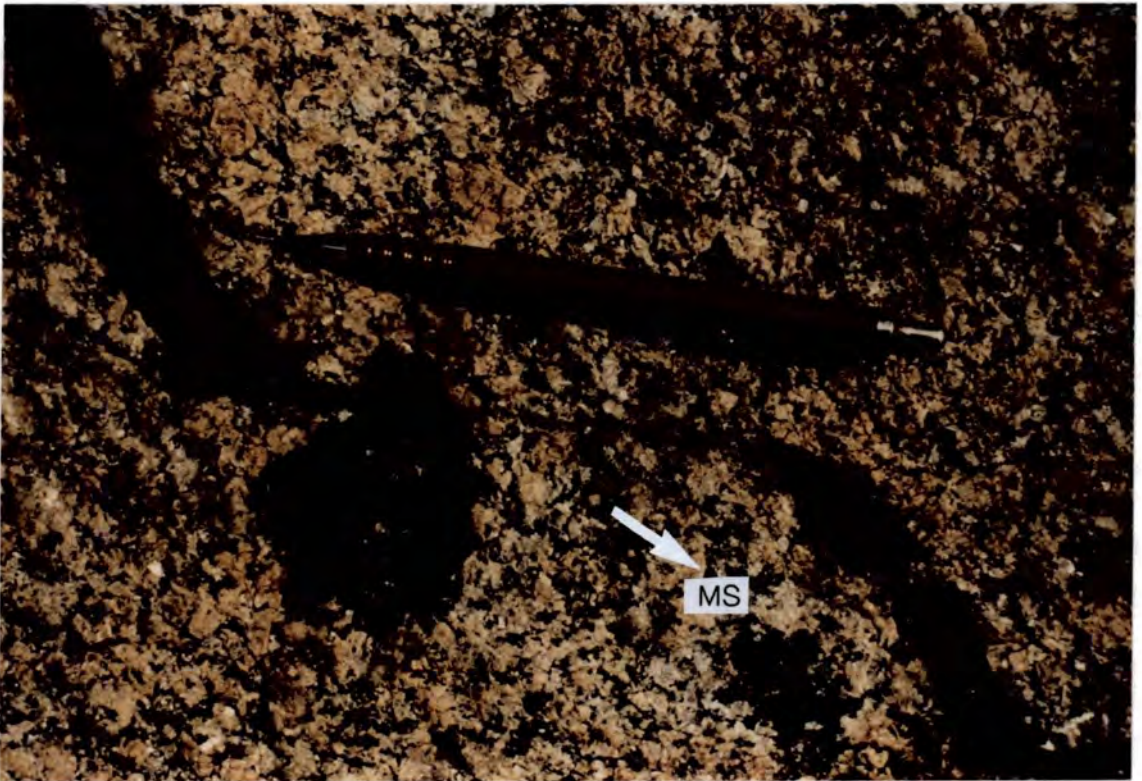


Plate 4.17 Sinistral shearing of a centimetre wide co-magmatic diorite dyke in an outcrop of Thorrr Granite exposed along the NW coast of Donegal. The narrow shear zone, with centimetre-scale offset, lies anticlockwise of the main magmatic state deformation fabric (white arrow marked MS) in the granite.

between the host granite and the dioritic rocks were initially liquid-liquid contacts that essentially remained immiscible, due, in part, to rheological contrasts (see Ch. 1).

Some centimetre wide diorite dykes and streaks show evidence of having been sheared synchronous with emplacement. The sense of displacement on such millimetre wide, apparently syn-magmatic shears is generally sinistral (Plate 4.17), and involves offsets of up to a few centimetres. This is significant, since, if these dioritic magmas are comagmatic with the granitic magmas, then the small shear zones within the dioritic dykes are evidence of sinistral shearing synchronous with the emplacement of the Thorr Pluton.

Oglethorpe (1987) suggested that these basic rocks were derived from rather primitive magmas, since they are rich in MgO, Cr, Ni and Sr. Although it is not within the scope of this study to comment with certainty on the derivation of the Thorr magma, it seems reasonable to suggest that the basic magmas seen along the north coast of Donegal may represent primitive and, possibly, parental magmas to the Thorr Granite. Such a proposition seems consistent with the view of Dempsey *et al.* (1990), who concluded, on the basis of Sm-Nd and Sr systematics, that the Thorr magma was a rather primitive magma, with little crustal input relative to some of the other Donegal granites.

4.3 *Fabrics within the Thorr Pluton*

The fabric evolution in a cooling pluton directly reflects the rheological behaviour of the magma in response to the imposed stresses that cause the deformation. Magma rheology is itself directly related to changes in the crystal to liquid ratio. The change from magmatic state to solid state deformation and the brittle-ductile transition mark the two main fundamental changes in the rheology of an igneous rock crystallised from a magma. In practice, both transitions often mark a change from homogeneous fabric development, typical within the magmatic interval, to heterogeneous strain distributions, often within narrower zones of shearing. Changes from one type of behaviour to another are intimately associated with increases in the bulk strength of both the magmatic and solid fractions as a result of decreases in the proportion of melt remaining in the system. Therefore the study of granite fabrics, in both the magmatic

and solid state, can provide much information about the thermal and mechanical state of a magma at the time of deformation (c.f. Gapais 1989).

Approximately 200 thin sections were examined in this study. The aim of the microscopic study was to determine the types and sequences of fabrics present in the pluton, and to comment on the orientation and kinematics of the fabrics where possible. These observations were used to deduce the thermal and mechanical evolution of the granite from magma to solid and, ultimately, to relate them to a mode of emplacement. It is not an aim of this thesis to comment on the specific deformation mechanisms that may have resulted in the development of textures within the granitic rocks studied. However, where relevant certain mechanisms will be alluded to.

4.3.1 *Magmatic state deformation fabrics*

These are characterised by a shape preferred orientation of euhedral phenocryst phases set in a groundmass of undeformed equant quartz (Ch. 1). The majority of deformation fabrics within the Thor Pluton are magmatic state fabrics (Ch. 1), indicating that deformation occurred early in the crystallisation history, mostly before the magma had reached its rheologically critical melt percentage (RCMP; Ch. 1). They are generally moderate- to well-developed fabrics, with a strong shape preferred orientation, which is often bimodal (see section 4.5). A typical magmatic state deformation fabric as seen in the field is illustrated in Plate 4.18. Microscopically, the magmatic state fabrics are clearly seen to be produced by alignment of euhedral plagioclase, euhedral to subhedral microcline microperthite and euhedral hornblende and biotite (Plate 4.19). Where mafic minerals are not abundant the magmatic state fabric is defined by alignment of feldspar phenocrysts only.

Magmatic state fabrics are generally well developed throughout the pluton, except in the granite forming the NE of the pluton, around Binanea Strand. Here randomly orientated igneous textures dominate (see section 4.1.2). They are typically equigranular and frequently exhibit graphic intergrowths of quartz and feldspar. The texture of the granite in this area is consistent with emplacement of this part of the granite at a slightly higher crustal level in a relatively passive manner.

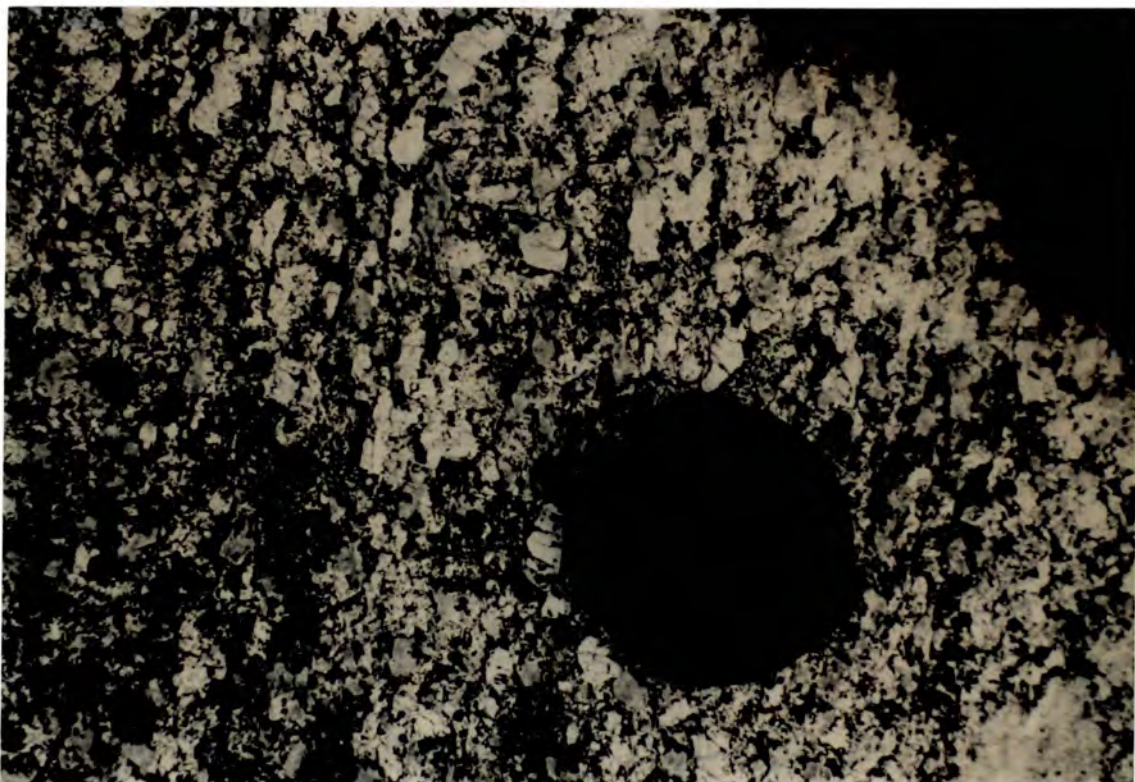


Plate 4.18 Typical magmatic state deformation fabric in an outcrop of Thorr Granite. The fabric is defined by a shape preferred orientation of feldspar and mafic minerals, quartz is generally interstitial.

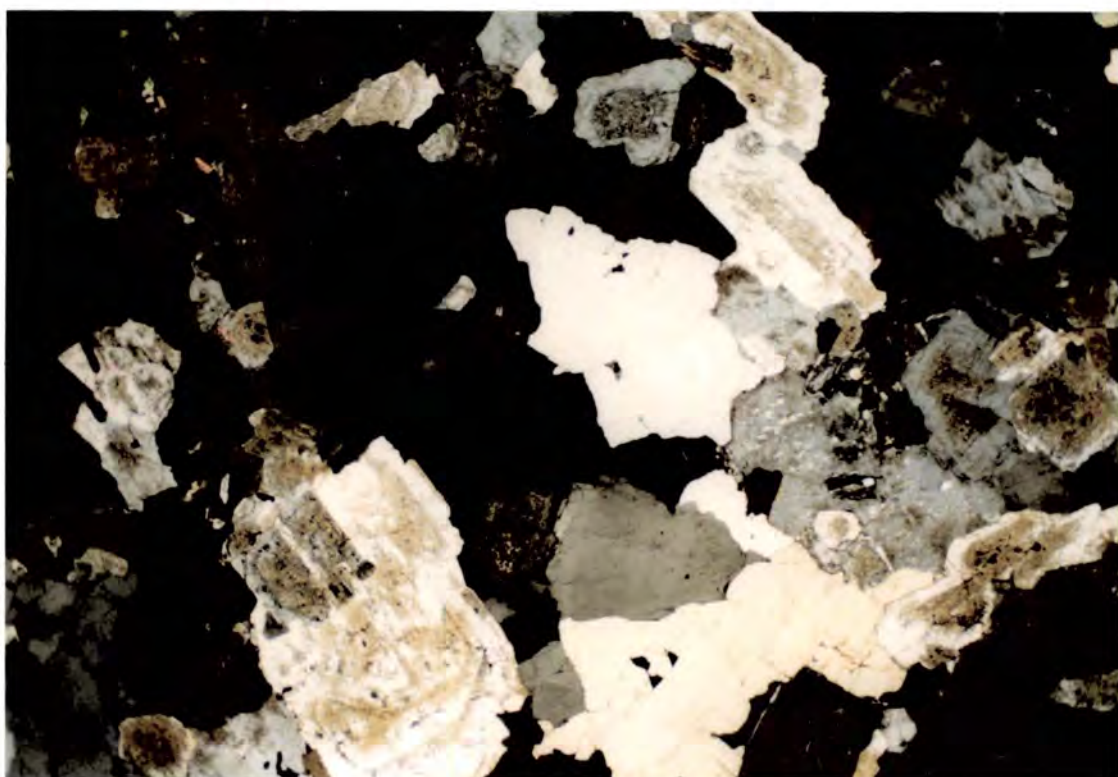


Plate 4.19 Photomicrograph of magmatic state deformation fabric in a specimen of Thorr Granite. The fabric in this specimen is predominantly defined by aligned, euhedral feldspar phenocrysts set in a groundmass of relatively equant quartz. (Field of view 18mm).

4.3.2 *Solid state deformation fabrics*

4.3.2.1 *General textural characteristics*

These are formed after a magma has reached its RCMP and coincide with an increase in bulk strength by several orders of magnitude (Ch. 1). The onset of solid state deformation is not marked by obliteration of the earlier magmatic state fabrics; indeed solid state deformation often initially enhances the preferred orientation of phenocryst phases developed during magmatic state deformation. The general characteristic is that all minerals begin to develop some degree of internal plastic deformation. Changes to the essential mineral phases in a granitic rock are as follows:

Feldspar: crystal shapes become modified by subgraining and recrystallisation of crystal margins, so that grains vary from euhedral to subhedral. This is particularly well illustrated in some specimens of the Thorr Granite, as a result of the primary subhedral nature of many of the alkali feldspars, since these subhedral projections are often the first areas to undergo recrystallisation processes. The magmatic state shape preferred orientation is, however, retained at this stage. Deformation lamellae are often formed within feldspars due to the accumulation of dislocations within the crystal structure. Alkali feldspars often show evidence of subsolidus exsolution to perthite and antiperthite, a process which may be enhanced by deformation (Simpson & Wintsch 1989). Plagioclase also experiences subsolidus alteration, especially small crystals, resulting in partial or total obliteration of oscillatory zoning and sometimes twinning. Twinning in feldspars may also be erased in areas where myrmekite is developed. The formation of myrmekite is basically a cation exchange process, but one which is enhanced by the build up of dislocations in areas of feldspar crystals that face the principal normal stress direction. At higher temperatures, the accumulation of dislocations may be relieved by recovery accommodated dislocation creep, which results in subgrain formation and rotation (Tullis & Yund 1985). However, recovery becomes less efficient with decreasing temperature and rapid diffusion of cations then results in the replacement of alkali feldspar by plagioclase and quartz in the form of myrmekite. This replacement initially occurs around crystal margins, or along boundaries such as twin planes, but may spread with increasing solid state deformation. Ultimately, the plagioclase and quartz intergrowths are further

deformed and may undergo recrystallisation due to dislocation creep, producing fine grained zones within which large strains may then accumulate by grain boundary sliding (McCaffrey 1989).

Hornblende: this tends to form a resistant fraction to solid state deformation, remaining relatively undeformed internally. Occasionally, however, crystals of hornblende may be subgrained, resulting in subhedral as opposed to euhedral shapes.

Biotite: this is often bent, fractured or kinked. Crystals and crystal aggregates of biotite tend to bend around phenocrysts of feldspar and, as a result, they often define the solid state fabric orientation.

Quartz: this tends to be the crystal phase that is most susceptible to solid state deformation. Crystals and crystal aggregates become flattened and elongate and may form polycrystalline lenses. Solid state deformation initially results in undulose extinction due to accumulation of dislocations within the lattice structure. Further deformation results in subgrain formation and with further rotation, ultimately to recrystallisation of quartz grains into elongate aggregates of small recrystallised grains. These lenses also define the solid state fabric orientation.

The transition from magmatic state to solid state deformation fabrics within a cooling pluton may be a straightforward continuous sequence, in which case the solid state fabrics are often coplanar with the earlier magmatic state fabrics. In some cases, however, there may be a hiatus in the fabric development as a result of a break in the deformation history, either because of a cessation in deformation, or as a result of a shifting deformation front, perhaps due to strain partitioning. Subsequent solid state deformation fabrics must then be examined to determine whether they are coplanar with the earlier fabrics. In cases where late solid state fabrics are not coplanar with the pre-existing fabrics, this may indicate a rotational strain or a change in the deformation regime. Generally, as temperature falls within a cooling pluton, solid state deformation will tend to be focused into narrower zones. Likewise, if a pluton is subjected to deformation *after* complete crystallisation and cooling, deformation will tend to be concentrated into discrete zones as opposed to being pervasively distributed. This is because in a low-grade regional metamorphic setting, like that seen in NW Donegal, the heat of deformation from an external source is unlikely to bring the granitoid to within half of its melting temperature, which approximates to the amount of energy required to produce a plastic mass capable of developing homogeneous fabrics (Gapais 1989).

4.3.2.2 *Types of solid state fabrics in the Thorr Pluton*

Three basic types of solid state deformation fabric are developed in the Thorr Pluton: (1) a high temperature to brittle-ductile, coplanar fabric; (2) sinistral S-C fabrics and mylonites that span the transition from ductile to brittle-ductile behaviour; (3) weak brittle-ductile fabrics that appear to have a dextral sense of shear in the field, but no obvious shear sense in thin section.

High temperature to brittle-ductile solid state fabrics: these are best developed near the pluton margins, particularly in the SE of the pluton in the Thorr area (Map 1) and in the contact and transitional facies rocks around country rock rafts. They are also developed at impersistent internal contacts between granite facies of slightly different composition or grain size, for example, the Lough Anure Sheet (Map 2). Plate 4.20 illustrates a typical high temperature solid state fabric in the contact facies close to the pluton margin at Thorr. The solid state overprint to the earlier magmatic fabric results in rounding of feldspar phenocrysts due to marginal subgraining. Some feldspar phenocrysts exhibit evidence of ductile internal lattice distortion in the form of undulose extinction. However, there is also evidence of brittle fracturing of feldspar, indicating deformation at temperatures $\leq 450^{\circ}\text{C}$ (Simpson 1985). Laths of biotite are bent around feldspar phenocrysts, but are mainly orientated parallel with the magmatic state shape preferred orientation. Groundmass quartz is flattened, subgrained and recrystallised. Individual subgrains within the aggregates exhibit strained extinction, indicating internal lattice distortion. These flattened quartz lenses are also subparallel to the earlier magmatic state fabric, indicating that this solid state deformation event was coaxial with the deformation that resulted in the production of the magmatic state fabrics.

Plate 4.21 is a photomicrograph of a rather intense, but relatively high temperature solid state fabric from a specimen of the Lough Anure Sheet. This solid state fabric is also coaxial with the magmatic state fabric. In this finer grained specimen there is evidence of quartz having undergone at least a small amount of grain boundary migration, resulting in the growth of new grains. This produces serrated grain boundaries in quartz and often results in the inclusion of small, aligned phyllosilicate minerals within new, relatively strain free quartz grains. Myrmekite is commonly developed at the edges of feldspar phenocrysts, but is prone to sericitisation. Hornblende and biotite are commonly altered to chlorite and epidote.

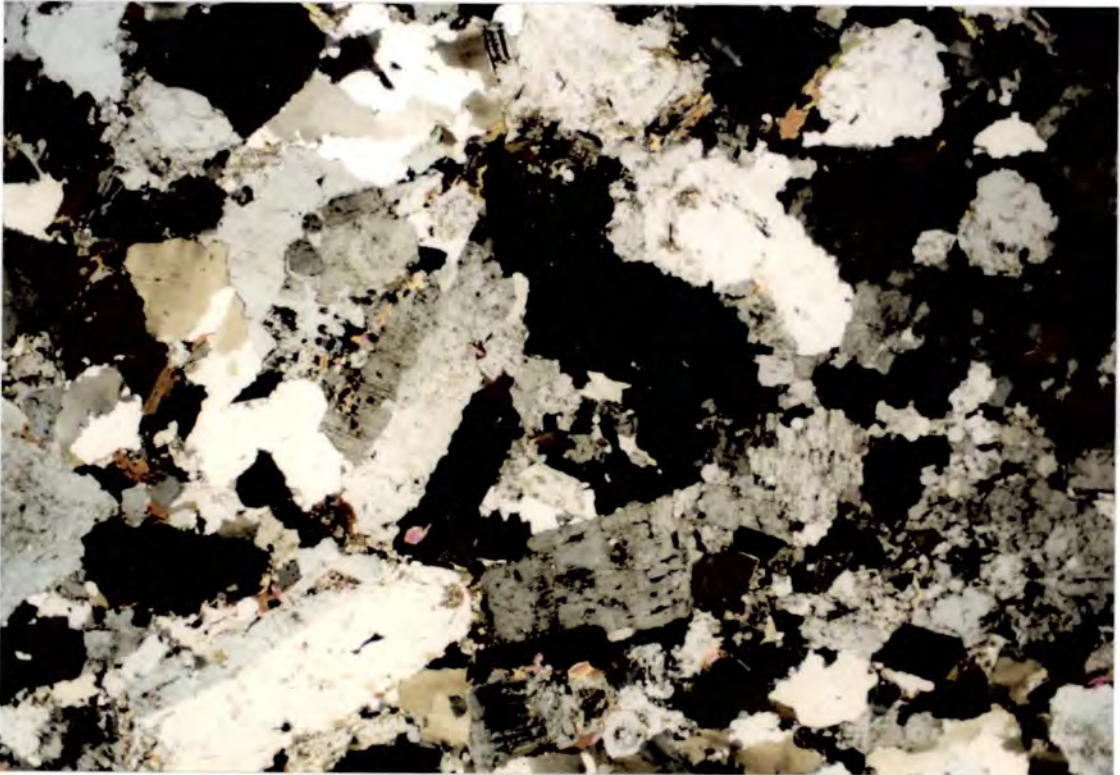


Plate 4.20 Photomicrograph of Thorr Granite that has suffered high temperature solid state deformation. The solid state fabric coaxially overprints the magmatic state fabric. It involves limited subgraining and recrystallisation around feldspar phenocrysts and groundmass quartz. (Field of view 18mm).



Plate 4.21 Photomicrograph of specimen of Lough Anure Sheet which carries a magmatic state deformation fabric overprinted by a high temperature solid state fabric. The solid state deformation results in recrystallisation of the margins of feldspar phenocrysts and, more pervasively, groundmass quartz. Note also the abundance of foxy red coloured biotite indicating that this is a 'contact facies' of the Thorr Granite. (Field of view 18mm).

The fact that solid state fabrics are developed in the contact facies rocks around rafts of country rocks, but not pervasively in the surrounding granite suggests that the raft-granite interfaces were sites of localised deformation, possibly as a result of the elevated water vapour pressures (Oglethorpe 1987) in the contact facies rocks.

Sinistral S-C fabrics and mylonites: these are mainly developed in the Crovehy Hills (Map 2) where the Thorr Granite is affected by deformation associated with the Main Donegal Granite Shear Zone (MDGSZ). Approximately 2km away from the MDGSZ magmatic state deformation fabrics in the Thorr Pluton are rotated from a NNW orientation to a WNW orientation (see section 4.4 and Map 2). Moving southwards towards the MDGSZ, the magmatic state foliation becomes overprinted by a solid state foliation. This is rather weakly developed at first, but it is nevertheless, easily recognised using the naked eye up to 1.5km away from the MDGSZ (Plate 4.22). It intensifies towards the shear zone, resulting in the development of a rather pervasive S-C fabric. Sense of shear is always sinistral, since C-planes are always anticlockwise of the magmatic state, S foliation (see stereonet in section 4.4). At ≤ 1 km away from the MDGSZ metre-scale mylonite zones are locally developed, which carry intense subhorizontal, ENE trending stretching lineations (Plate 4.23). This progressive localisation of deformation is a typical product of active deformation in a cooling pluton (Gapais 1989).

Microscopically these fabrics can be seen to intensify towards the MDGSZ. At approximately 1.5km away from the shear zone they are characterised by subrounded feldspar phenocrysts with pressure shadows of quartz and phyllosilicates, that give a sinistral sense of shear (Plate 4.24). Quartz has the form of polycrystalline lenses within these pressure shadows, and more generally within the groundmass. Phyllosilicates are bent around the feldspar phenocrysts as well as being included within pressure shadows (Plate 4.24). As the fabrics intensify and become mylonitic, grain size is generally reduced and feldspar tends to suffer more brittle deformation

Intense solid state fabrics, similar to those described above, are seen deforming the Thorr Granite at Cleengort Hill (Map 2). The granite here is typically finer grained than at Thorr, it is possible that this is, at least partly, due to grain size reduction in response to solid state deformation. There is evidence of brittle-ductile deformation here also, since feldspar tends to be brittly fractured, whereas quartz tends to be



Plate 4.22 Pervasive S-C fabric in the Thorr Granite exposed in the Crovehy Hills. This is typical of the Thorr Granite in a 1-2km wide, ENE - trending belt that has been affected by movements on the MDGSZ.



Plate 4.23 Intense subhorizontal, ENE - trending stretching lineation exposed in protomylonitic Thorr Granite close to the trace of the MDGSZ in the Crovehy Hills.



Plate 4.24 Photomicrograph of sinistral shear sense indicators in protomylonitic Thorr Granite. These typically consist of mica fish or porphyroclasts of feldspar surrounded by quartz - mica pressure shadows. (Field of view 4.4mm).

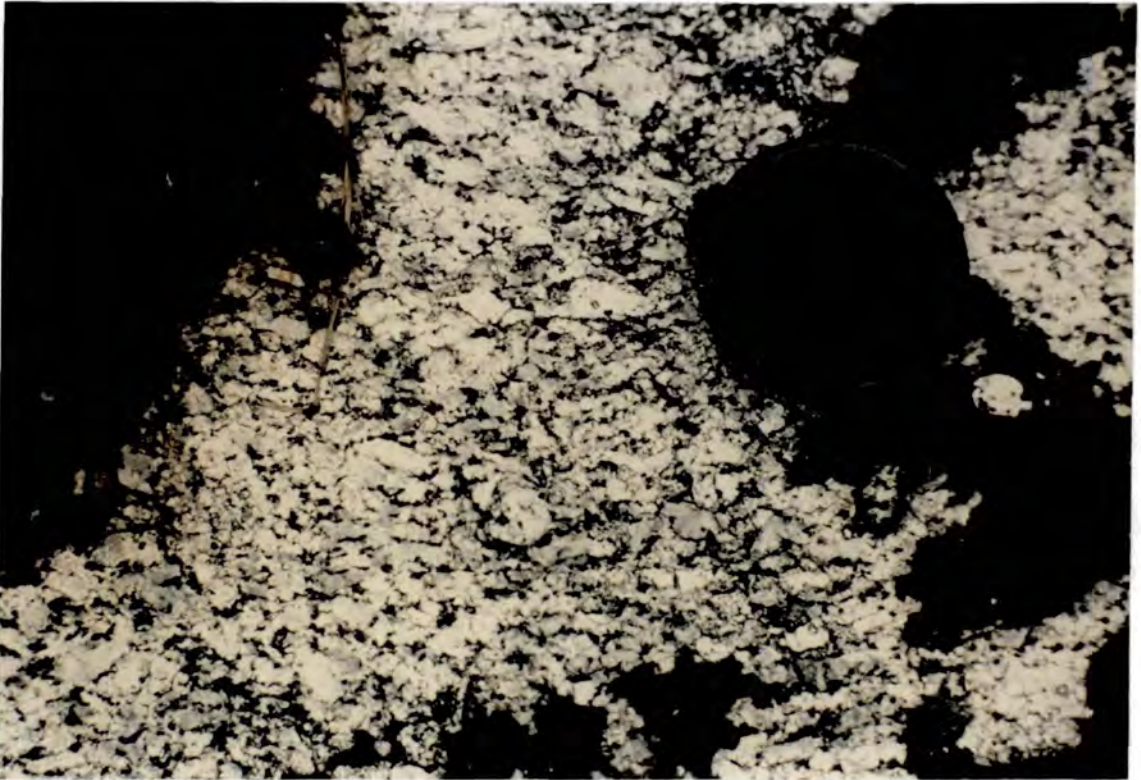


Plate 4.25 Weak brittle-ductile solid state deformation fabric in the Thorr Granite exposed inland and to the SE of Maghery.

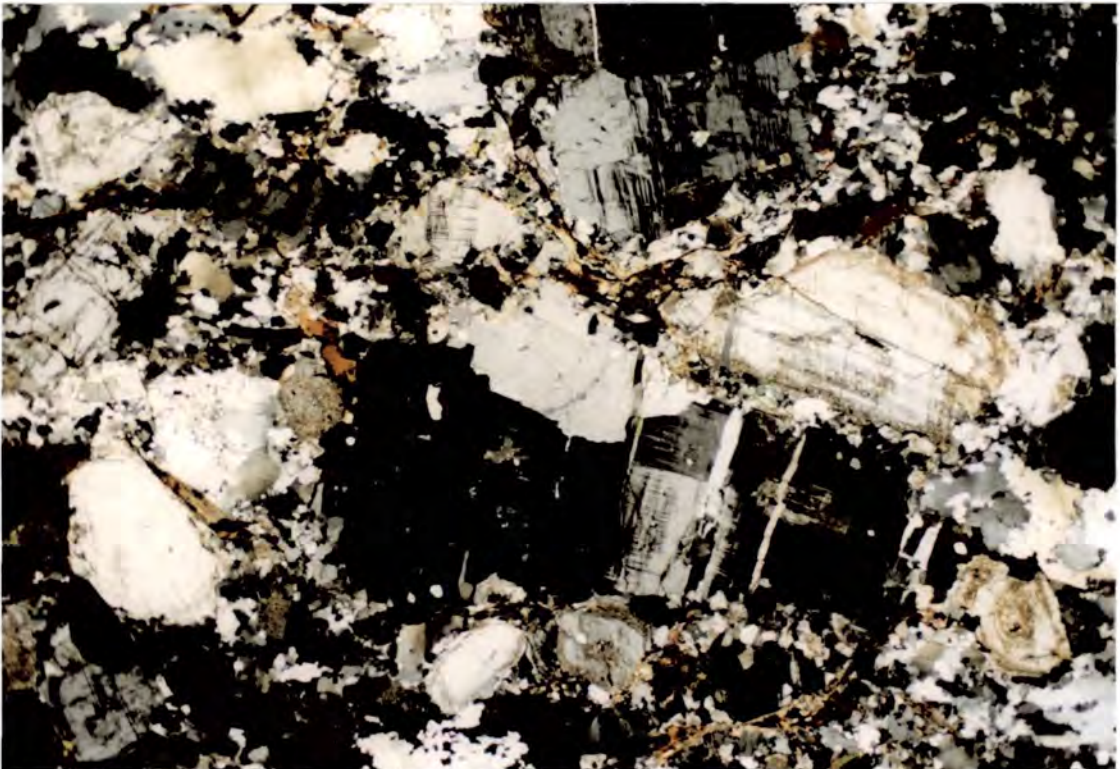


Plate 4.26 Photomicrograph of specimen of Thorr Granite collected inland and SE of Maghery. The specimen illustrates the occurrence of a weakly developed brittle-ductile solid state deformation fabric. Note the occurrence of sinuous brittle fractures around the margins of and sometimes through feldspar phenocrysts. These are filled with fine-grained phyllosilicates and ore minerals. The phenocrysts have been eroded by solid state deformation so that they are no longer euhedral. Note also that groundmass quartz has deformed generally by ductile processes such as recovery and recrystallisation. (Field of view 18mm).

deformed by internal lattice distortion. This is again interpreted as being a product of shearing along the MDGSZ, the main trace of which is located ENE of Cleengort Hill.

Weak brittle-ductile solid state fabrics: these are observed in the granite inland from Maghery in a 500- to 750m wide zone (Map 2). In the field they are observed due to a small amount of flattening of quartz in particular. Flattened aggregates of quartz sometimes appear to form pressure shadows around subhedral feldspar phenocrysts that, in the field, appear to have a top to the right, or dextral sense of shear, but which does not photograph particularly well (Plate 4.25). Sense of shear could not be determined with certainty in thin section. The texture of this fabric in thin section is dominated by a general reduction in grain size which is achieved by ductile processes, such as subgraining and recrystallisation, in quartz and biotite, but by rather more brittle mechanisms, such as fracturing, in feldspar (Plate 4.26). There is no evidence of substantial secondary recrystallisation by grain growth in these rocks.

The textural features of this fabric indicate that it is a rather low grade feature. This is consistent with the texture seen in low grade dextral shear bands observed within the country rocks next to the granite contact north of Maghery Bay (Ch. 3), suggesting this episode of dextral shear post-dates the emplacement and much of the cooling history of the Thorr Granite.

4.4 The geometry and kinematics of the magmatic state deformation fabrics

Figure 4.2 summarises all data relating to magmatic state foliations and lineations collected during this study of the Thorr Pluton. Throughout the pluton the magmatic state foliation is much better developed than the mineral lineation, meaning that the Thorr Granite is an $S > L$ tectonite (Flinn 1965). In most areas two subfabrics, A & B, can be defined. Stereonets (a) and (b) relate to subfabric A, which is defined by feldspars with relatively low axial ratios, and stereonets (c) and (d) relate to subfabric B, which is defined by feldspar phenocrysts and mafic minerals with relatively high axial ratios. The contoured plots of poles to foliation (nets a & c) illustrate that both subfabrics have a steep to subvertical dip. The strike of the foliation is variable, reflecting the swing in the fabrics around the pluton. However, density

TOTAL FABRIC DATA

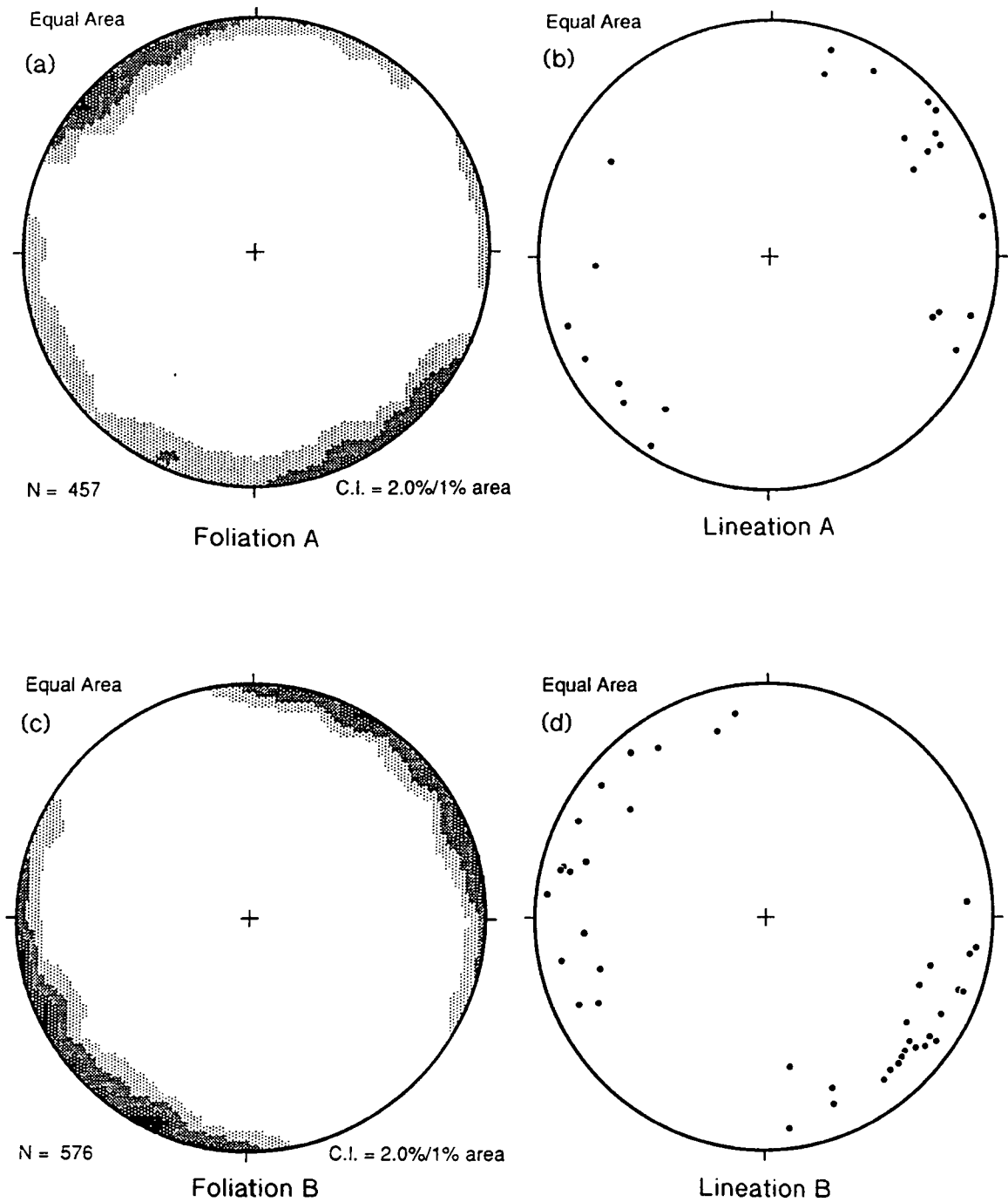


Figure 4.2 Equal area stereonet of magmatic state foliation and lineation data from around the pluton. (a) Contoured plot of poles to foliation A (low axial ratio); (b) mineral lineations in the plane of foliation A; (c) contoured plot of poles to foliation B (high axial ratio); (d) mineral lineations in the plane of foliation B.

maxima indicate that on average, subfabric A has a NE-ENE strike and subfabric B has an ESE-SE strike. Mineral lineations relating to both subfabrics are shallowly plunging to subhorizontal.

Unimodal fabrics occur where feldspar crystals have approximately equal axial ratios. Where observed (mainly along the NW and N coasts) they are generally defined by feldspar laths with relatively large axial ratios and by mafic minerals. They have been included with subfabric B in the appropriate subareas on the basis of their axial ratios.

Figures 4.3 to 4.13 summarise foliation data from around the pluton according to subareas outlined in Figure 4.1. Within these subareas the magmatic state foliation is generally steeply orientated and mineral lineations, where observed are generally shallowly to moderately plunging. In the Thorr area, next to the main contact with the mobilised country rocks, N-NW (see Map 2), steeply dipping magmatic state deformation fabrics and rare subhorizontal, N-S trending mineral lineations are interpreted as being consistent with shearing along a steeply dipping, N-S orientated shear zone. Evidence from the country rocks in this area (Ch. 3) indicates that the sense of movement on this shear zone during granite emplacement was sinistral.

Subfabrics show a consistent sense of obliquity, with subfabric A (low axial ratios) generally striking clockwise of subfabric B (high axial ratios). This obliquity can be used to determine a sense of shear in relation to deformation in the magmatic state (see section 4.5).

In the Crovehy Hills, magmatic state deformation fabrics consistently strike WNW (Fig. 4.13a) as opposed to the N-striking fabrics seen in the Thorr area (Fig. 4.12a). This swing in the strike of the magmatic state fabrics is believed to be related to sinistral shearing along the MDGSZ which results in reorientation of the magmatic fabrics to the WNW trend. Solid state deformation fabrics in this area (Fig. 4.13c) consistently strike ENE, an orientation which is anticlockwise of the magmatic state fabrics. This is also consistent with sinistral shearing and is believed to be associated with movements on the MDGSZ. The evolution of the deformation fabrics in the Crovehy Hills is consistent with initiation of sinistral shearing along the MDGSZ *before* the Thorr Granite had completely crystallised.

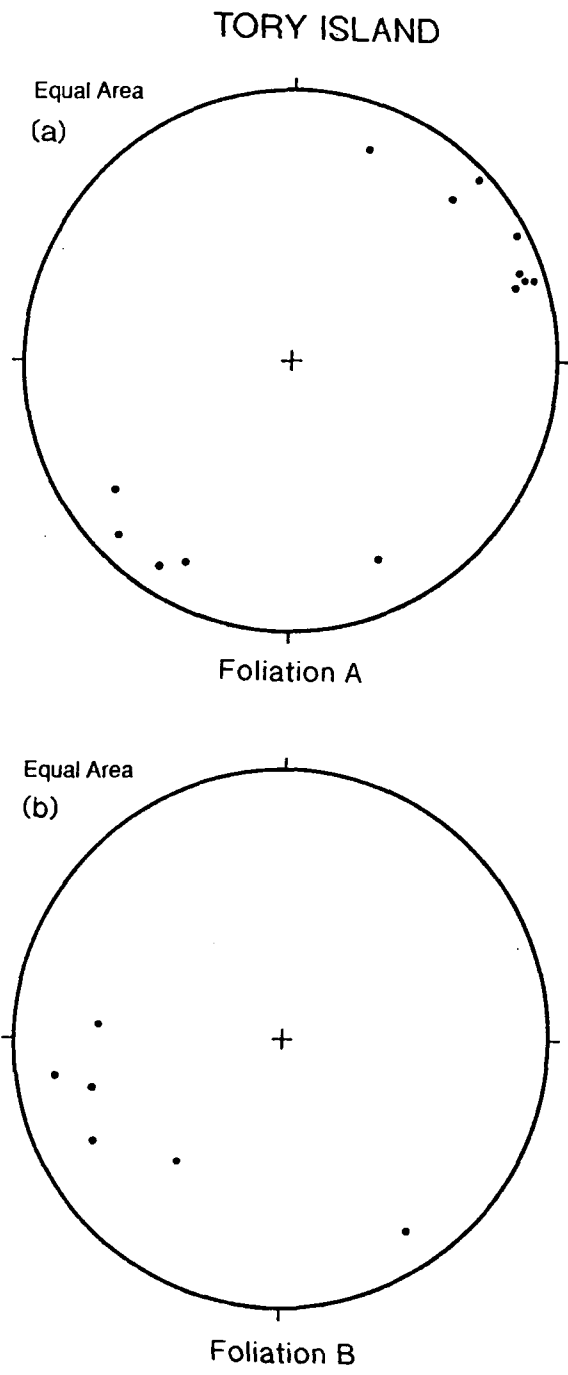
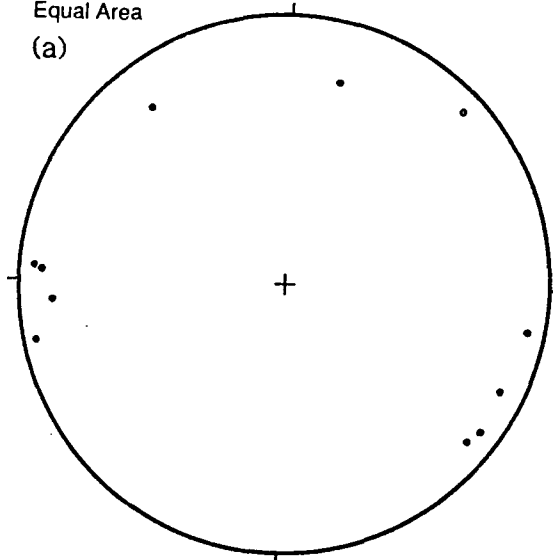


Figure 4.3 Equal area stereonet of magmatic state fabric elements in subarea 1. (a) Poles to foliation A; (b) poles to foliation B.

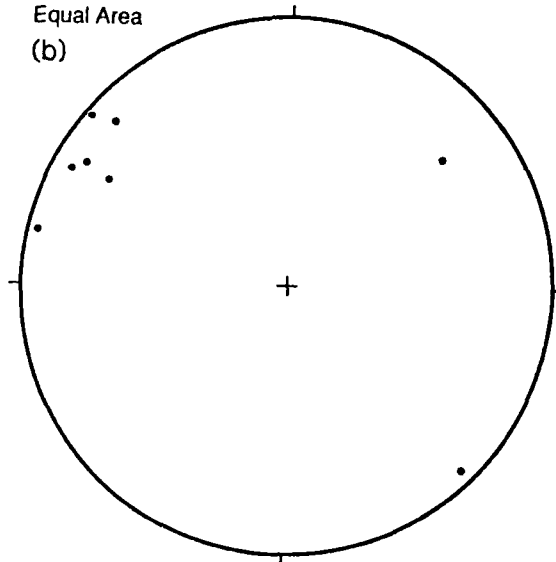
INISHBOFIN ISLAND

Equal Area
(a)



Foliation A

Equal Area
(b)



Foliation B

Figure 4.4 Equal area stereonet of magmatic state fabric elements in subarea 2. (a) Poles to foliation A; (b) poles to foliation B.

NORTH COAST

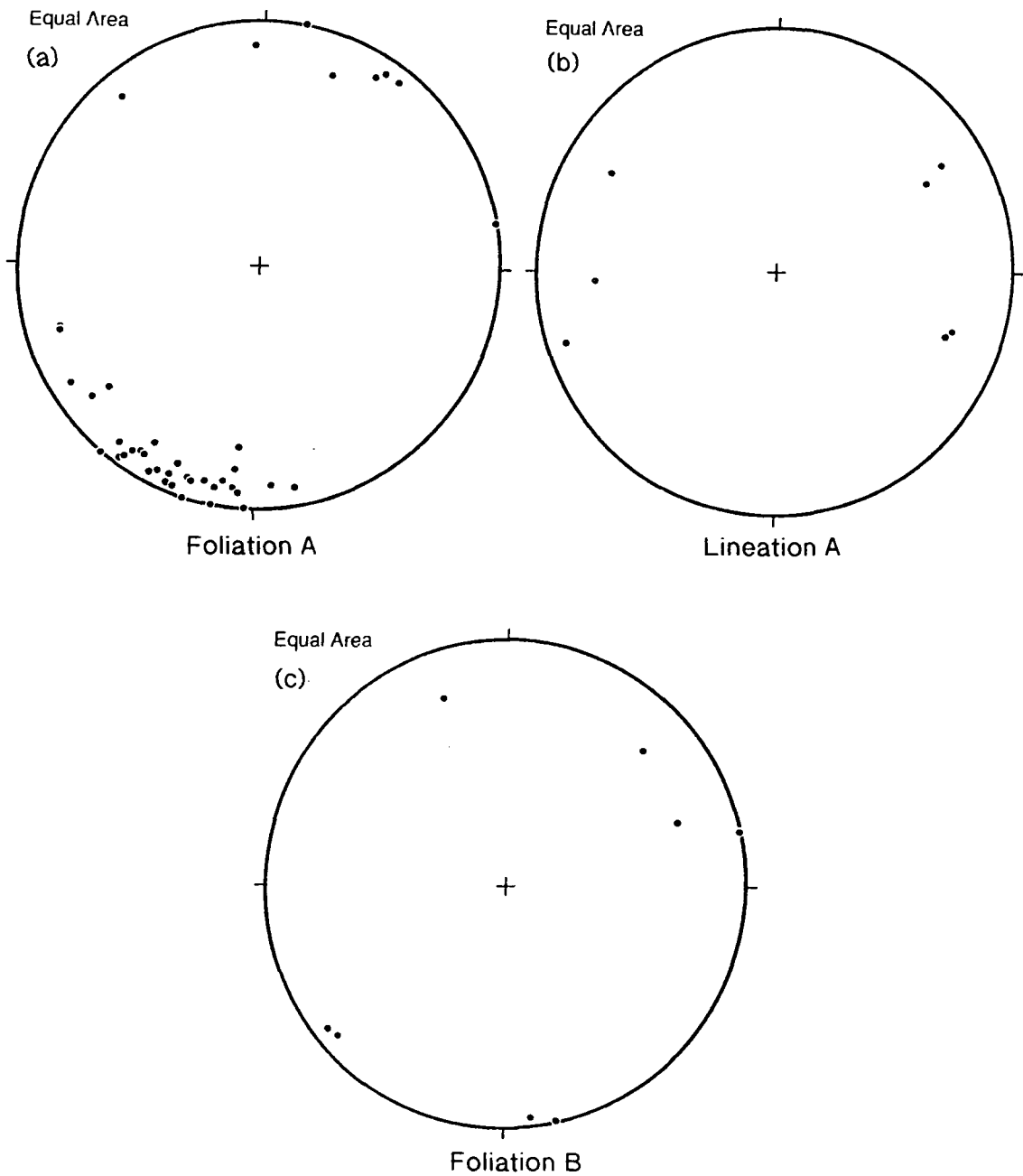


Figure 4.5 Equal area stereonet of magmatic state fabric elements in subarea 3. (a) Poles to foliation A; (b) mineral lineation in foliation plane A; (c) poles to foliation B.

NORTHWEST COAST

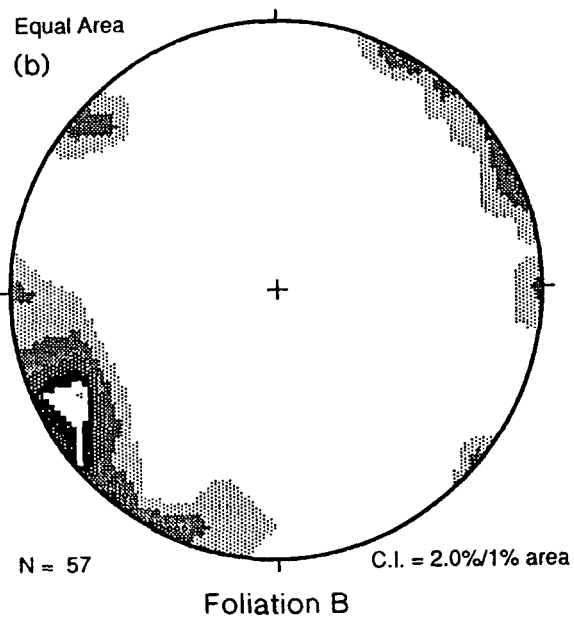
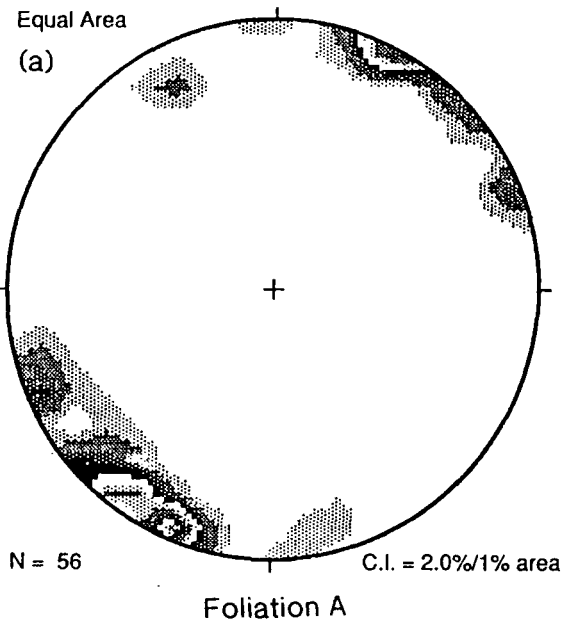


Figure 4.6 Equal area stereonets of magmatic state fabric elements in subarea 4. (a) Contoured plot of poles to foliation A; (b) contoured plot of poles to foliation B.

WEST COAST

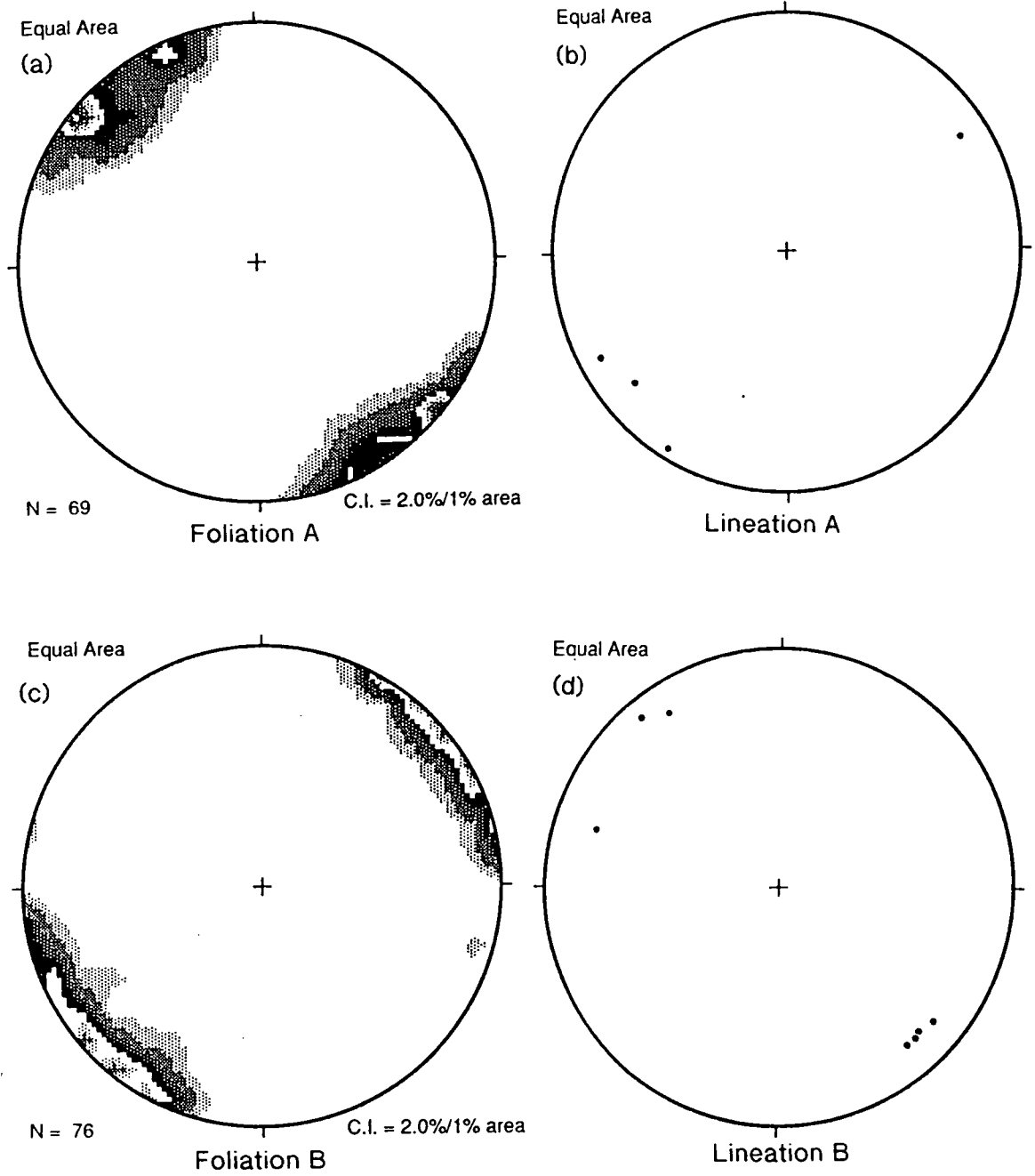


Figure 4.7 Equal area stereonets of magmatic state fabric elements in subarea 5. (a) Contoured plot of poles to foliation A; (b) mineral lineations within foliation plane A; (c) contoured plot of poles to foliation B; (d) mineral lineations within foliation plane B.

CENTRAL INLAND

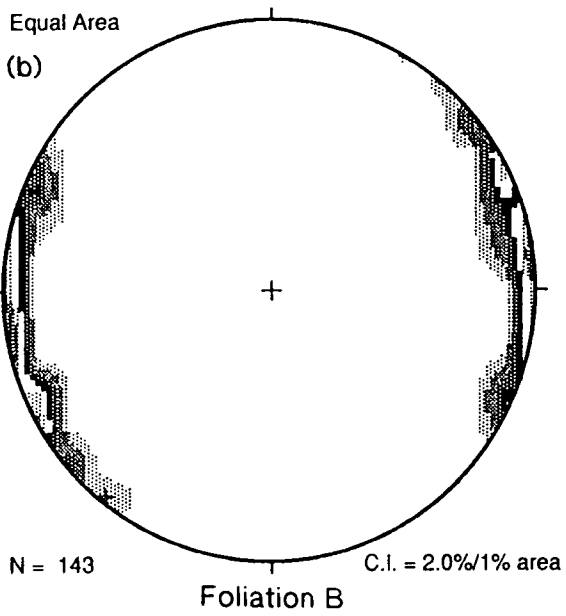
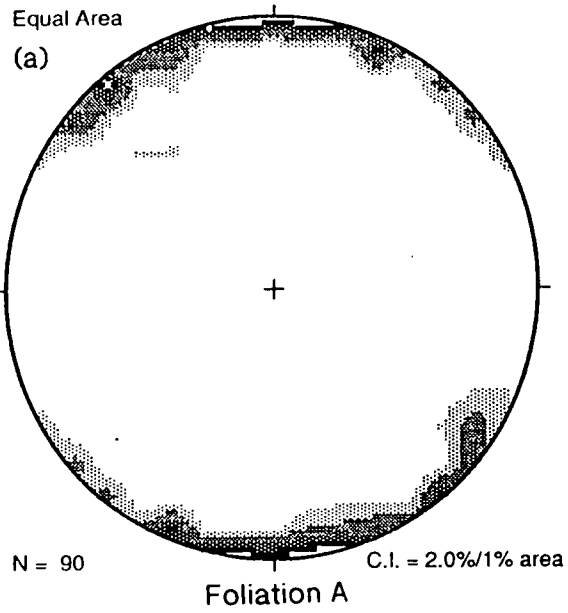


Figure 4.8 Equal area stereonets of magmatic state fabric elements in subarea 7. (a) Contoured plot of poles to foliation A; (b) contoured plot of poles to foliation B.

SOUTHWEST COAST

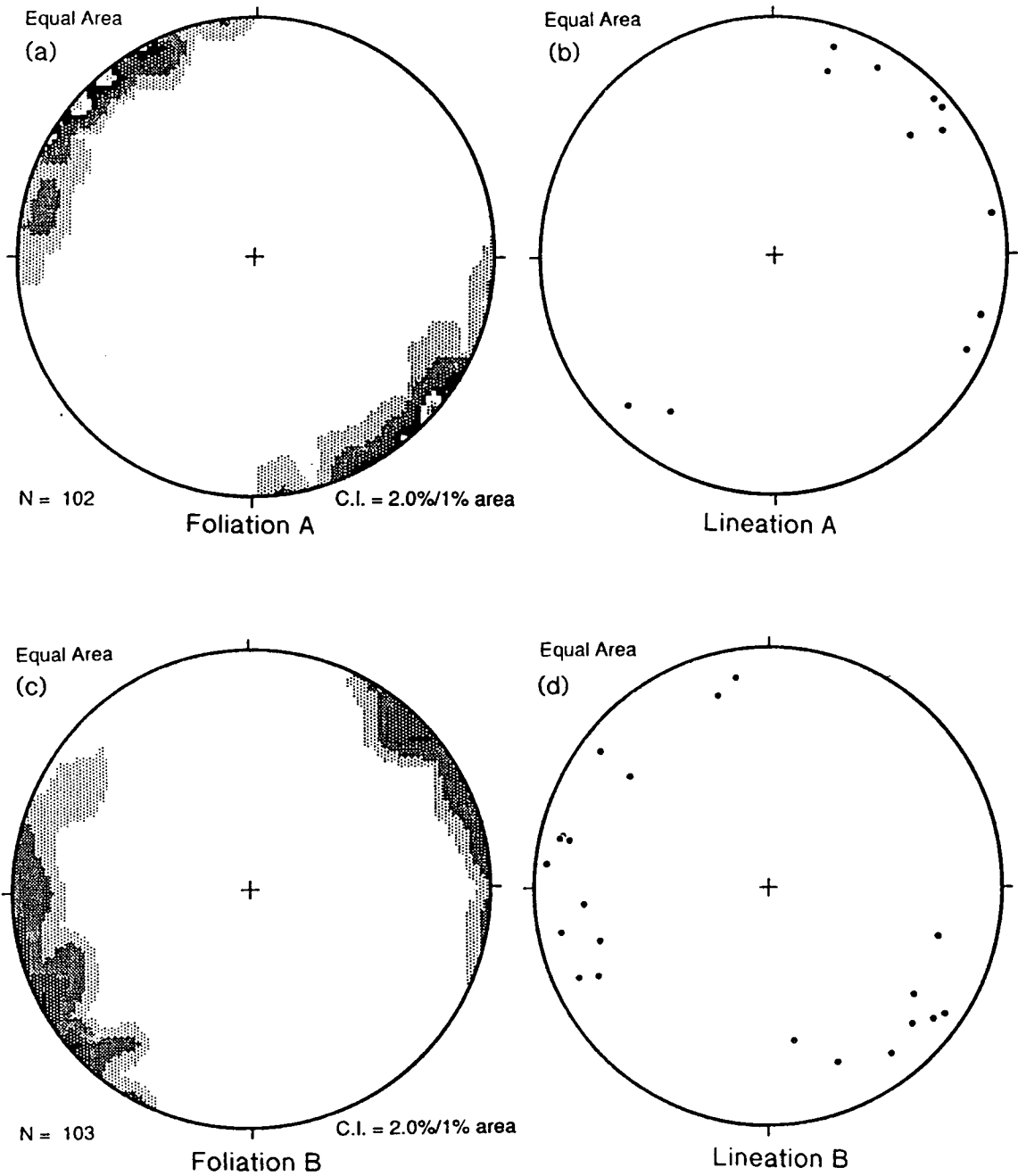


Figure 4.9 Equal area stereonets of magmatic state fabric elements in subarea 6. (a) Contoured plot of poles to foliation A; (b) mineral lineations within foliation plane A; (c) contoured plot of poles to foliation B; (d) mineral lineations within foliation plane B.

MAGHERY

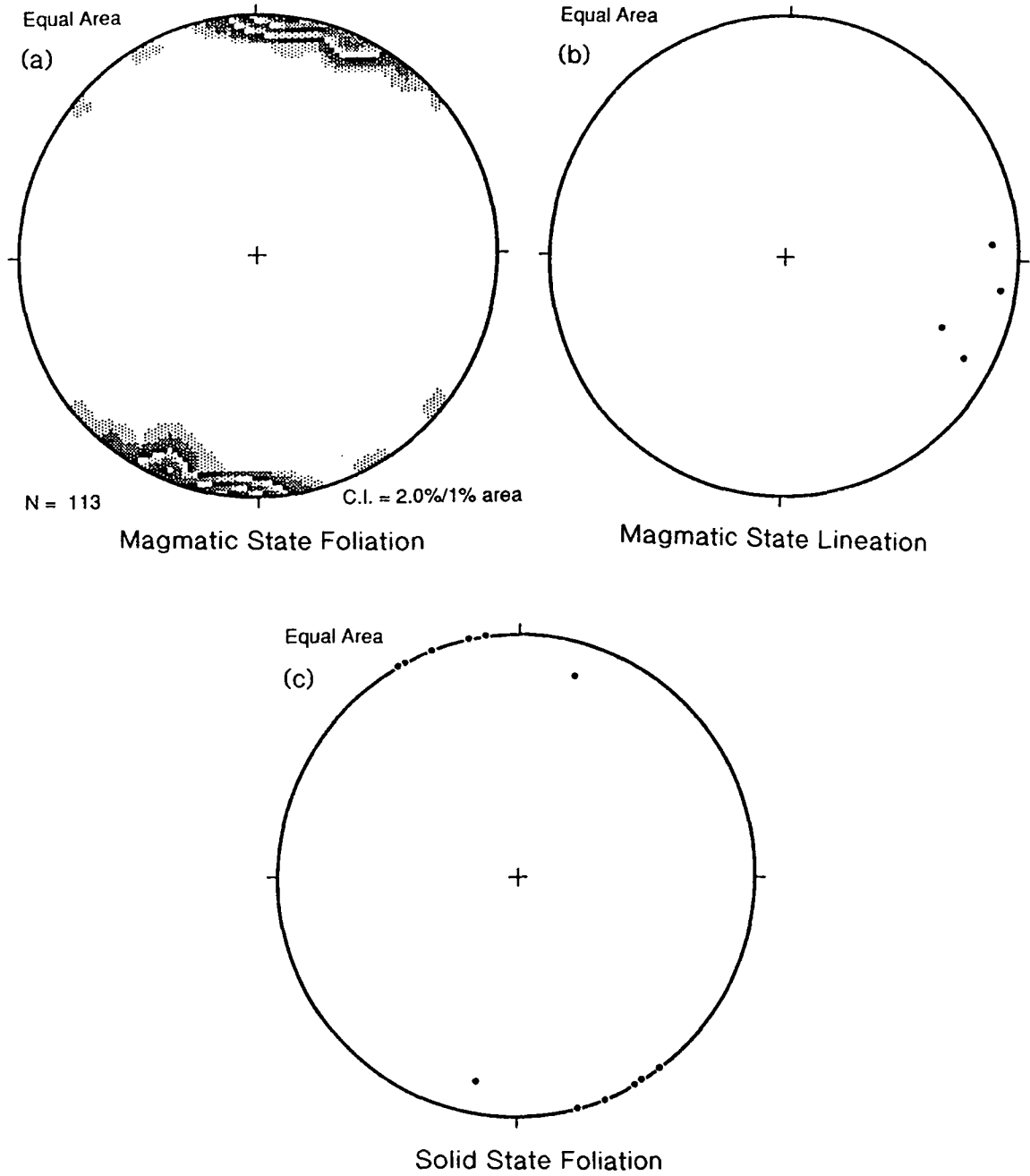


Figure 4.10 Equal area stereonets of deformation fabric elements in subarea 9. (a) Contoured plot of poles to magmatic state foliation; (b) mineral lineations within the magmatic state foliation plane; (c) poles to solid state fabric.

ARANMORE ISLAND

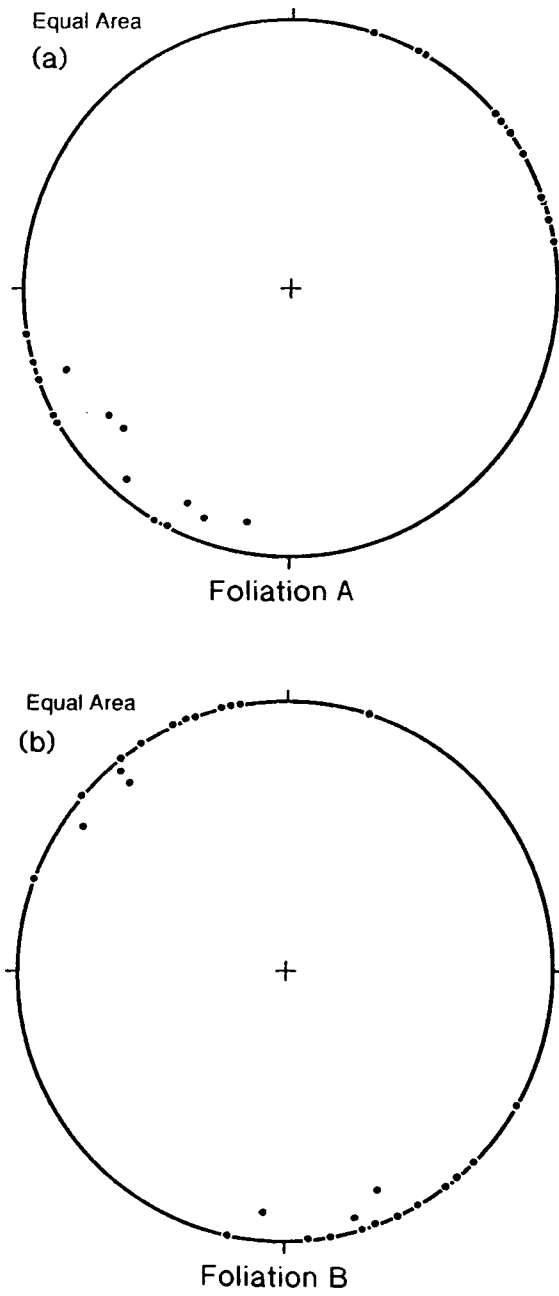


Figure 4.11 Equal area stereonets of magmatic state fabric elements in subarea 8. (a) Poles to foliation A; (b) poles to foliation.

MEENCORWICK - LOUGH AGHER - ARDVEEN

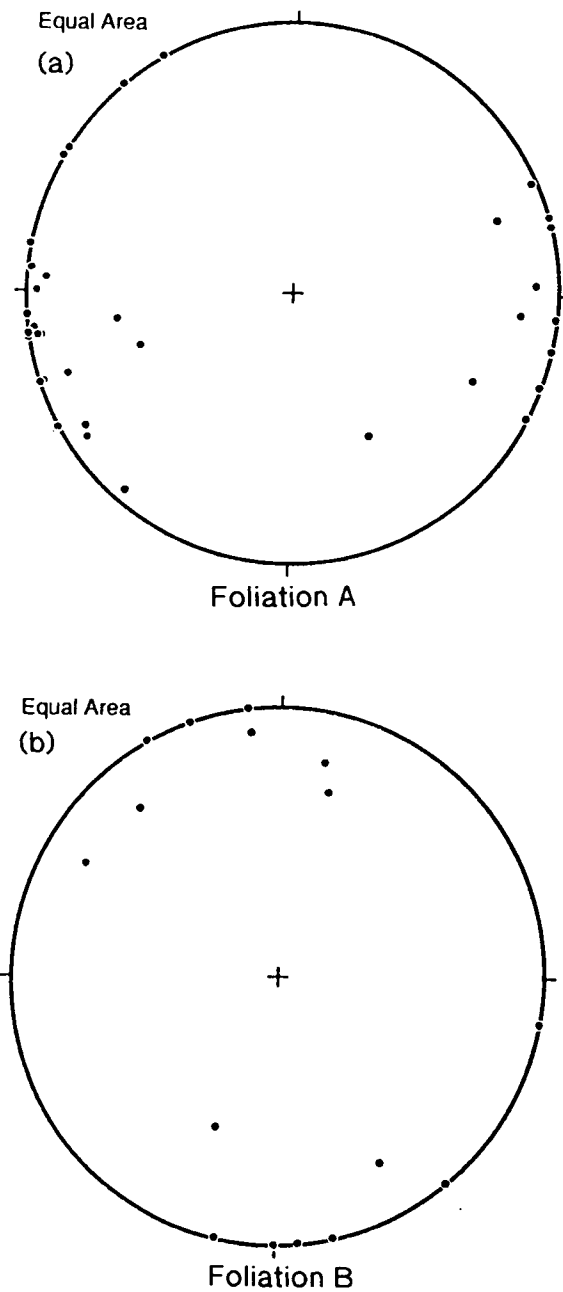


Figure 4.12 Equal area stereonets of magmatic state fabric elements in subarea 10. (a) poles to foliation A; (b) poles to foliation B.

CROVEHY HILLS

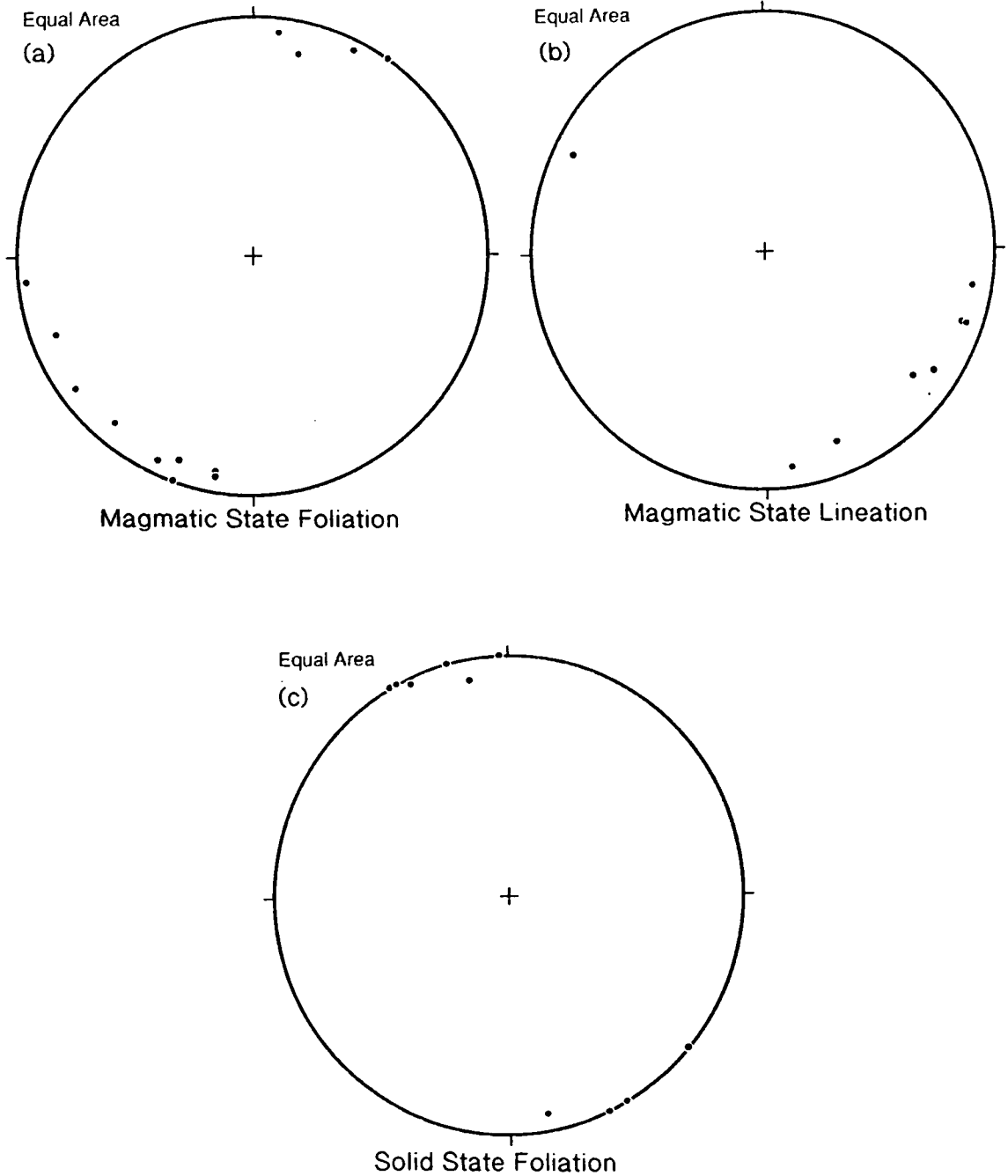


Figure 4.13 Equal area stereonets of deformation fabric elements in subarea 11. (a) Poles to magmatic state foliation; (b) mineral lineation in magmatic state foliation plane; (c) poles to solid state foliation.

4.5 Sense of shear indicators in the Thorr Granite

Blumenfeld and Bouchez (1988) summarised several criteria for determining sense of shear in deformed igneous rocks (see Ch. 1). The criteria used depends on whether the rock has been deformed in the magmatic state or the solid state. Deformation fabrics in the Thorr Granite mainly developed in the magmatic state (see section 4.3), hence potential criteria for determining sense of shear are:

- (1) obliquity between intrusion walls and a planar foliation defined by early formed phenocrysts;
- (2) sense of tiling of megacrysts (requires intermediate to high megacryst content);
- (3) obliquity between subfabrics (either shape dimensional or crystallographic).

The first of these methods is unsuitable in most areas, except in the southeast of the pluton, in the area around Thorr, since this is one of the few areas where the contact is clearly exposed. The obliquity between the intrusion walls and the magmatic state deformation fabric in this area is consistent with the fabric having formed during sinistral shearing. The second method could not be used since tiling of megacrysts does not appear to be a common phenomenon in the pluton. The author assumes that this is due to low concentrations and/or wide dispersion of megacrysts during deformation in the magmatic state, in agreement with Oglethorpe (1987). So, the latter of the three methods proved to be the most useful when examining the Thorr Pluton.

4.5.1 Obliquity between subfabrics as a method to deduce sense of shear

It is widely recognised that if there is a competence difference between a particle and its matrix during deformational flow, then the particle will undergo a rotary motion which is not the same as that of the matrix (see Ramsay 1967 and references therein). The motion of rigid particles contained in fluids undergoing slow lamellar flow (simple shear) has been investigated by many workers (Jeffery 1923; Gay 1968; Willis 1977; Fernandez 1982, 1984; Fernandez *et al.* 1983; Fernandez & Laporte 1991; Ildfonse *et al.* 1992). Jeffery (1923) found that the forces which acted on the particle could be reduced into two couples, one tending to make the ellipsoid adopt the same rotation as the surrounding fluid, and another

tending to set the ellipsoid so that its axes moved towards those of the principal distortion axes in the fluid (see Fig. 4.14). Jeffery also found that the equations defining the motion made by the rigid particle could be readily solved if the particle was either a uniaxial prolate or uniaxial oblate ellipsoid.

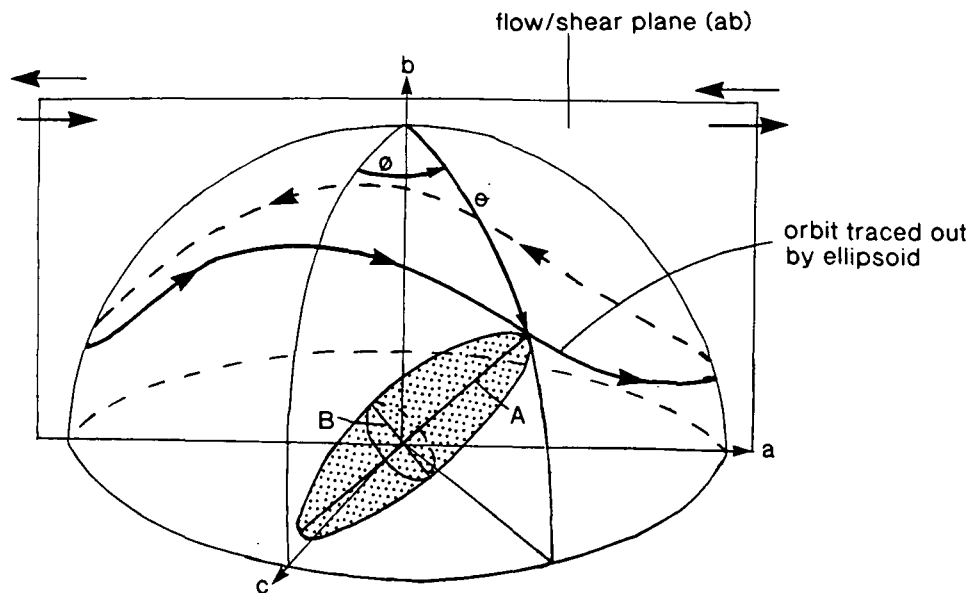


Figure 4.14 Sketch diagram to illustrate the orbit traced out by the end of the unique axis of a rigid uniaxial ellipsoid in a fluid undergoing lamellar flow, flow plane ab and flow direction a (after Ramsay 1967).

Flow of a fluid containing rigid particles leads to the development of preferred orientations of the rigid particles. The dimensions of the preferred orientation depends on whether the particles are uniaxial prolate or uniaxial oblate. Prolate particles tend to form linear fabrics, with the longest axes of the particles concentrated around the flow/shear direction (' a ' in Fig. 4.14). Oblate particles tend to form planar fabrics, since the minimum energy configuration is one where the short axis of the particle is at high angles to the shear plane and the longest axes are subparallel to the shear plane (see Ramsay 1967 for detailed discussion). In active transcurrent shearing in a sinistral sense, the b axis in figure 4.14 is equivalent to the Y axis of the incremental and the finite strain ellipse, and so oblate particles will rotate about this axis in response to regional deformation. The angular velocity of the particles about this axis is intimately related to the period of rotation about their own axes (Willis 1977; Fernandez & Laporte 1991). This in turn is affected by the aspect ratio of the particles; for example, rigid particles that have low aspect ratios have a smaller period of rotation than those with high aspect ratios. In the case of oblate particles, this means that those with low aspect ratios have a higher angular velocity, and, therefore, rotate into

parallelism with the shear plane faster than those with higher aspect ratios. If one considers a population of phenocrysts of different shapes, and hence, aspect ratios, suspended within a magma that is being actively deformed, the different rates of rotation of the phenocrysts will result in the development of subfabrics that are oblique to one another (Fernandez & Laporte 1991 and references therein).

Studies of deformed materials often reveals that a range of particle shapes is not uncommon, and the presence of subfabrics has been increasingly recognised over the past few years, particularly in igneous rocks. This has resulted in a re-emergence of interest as to what these fabrics can tell us about the deformation that resulted in their formation. Studies around this subject culminated in a number of key papers during the past two years (Fernandez & Laporte 1991; Ildefonse *et al.* 1992), the former of these works described a technique for determining shear sense from rocks deformed in the magmatic state, using obliquity between subfabrics.

Fernandez and Laporte (1991) stated that the preferred orientations displayed by magmatic rocks usually result from the rigid rotation of phenocrysts suspended in their slowly deforming ductile matrix (as outlined above), whatever the origin of this deformation (i.e. whether it is due to magmatic emplacement, regional stress field, etc. Fernandez 1984). In this context, contributing factors to the development of the final fabric are (a) the shape of crystals (or "markers"), (b) their initial distribution, (c) the deformation history. Assuming an initially isotropic distribution of rigid particles, Curie's Principle, which simply states that the symmetry of the factors contributing to a deformation cannot be higher than the symmetry of the deformed fabric (see Fernandez & Laporte 1991 for full explanation), implies that low symmetry Shape Preferred orientations (S.P.O., see Ch. 1) are developed in response to strain regimes, or polyphase strain histories, with a non-coaxial component.

If we consider the case of transcurrent simple shear, which seems to be the most appropriate strain regime that may be associated with the emplacement of the Thorr Pluton (based on contact relationships and kinematics), the orientation of the principal strain axes and planes of the finite strain ellipsoid have a geometric arrangement identical to that shown in figure 4.14. In this case, the XY plane is vertical, as is the Y axis of the finite strain ellipsoid, the XZ plane is, therefore, horizontal (see Fig. 4.15). In the case of oblate particles in such a regime, the fabric will develop by rotation of the particles

about the Y axis, with an angular velocity that is dictated by their axial ratio. In order to visualise the asymmetry of subfabrics developed in such a regime, one must examine the subfabrics in the XZ plane of the finite strain ellipsoid (Fernandez & Laporte 1991).

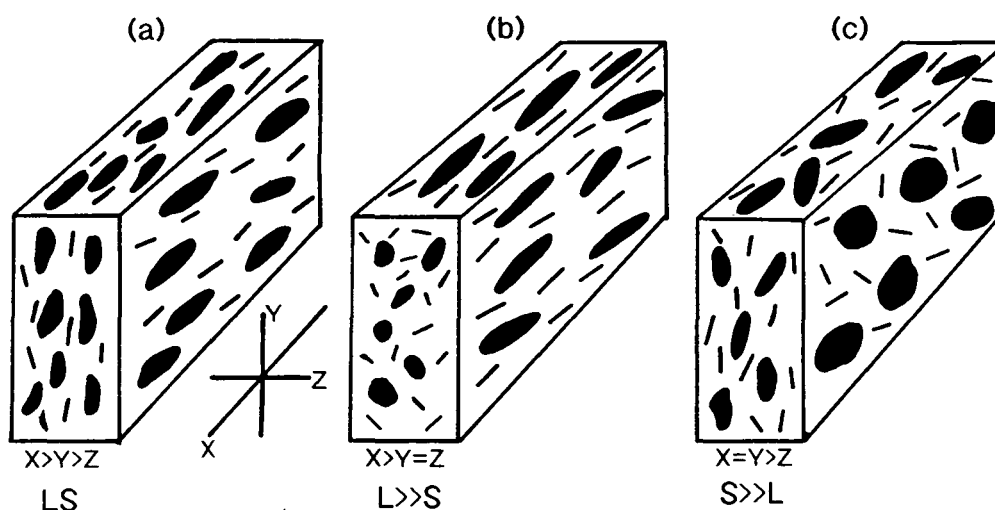


Figure 4.15 Sketch diagram to illustrate the orientation of deformation fabric elements in LS, L & S tectonites under a strain regime where X & Z are horizontally directed and Y is vertically directed (adapted after Hutton 1988).

Subfabrics in igneous rocks can be defined by different minerals, for example, biotite and feldspar, or by families of the same mineral with different axial ratios, for example feldspars. In the case of biotite and feldspar, the two mineral species define quasi-homogeneous populations of oblate spheroids, biotite being more oblate than feldspar. Fernandez and Laporte (1991) stated that simple shear acting upon such a heterogeneous population of axial markers produces monoclinic fabrics with a symmetry plane that coincides with the XZ deformation plane. The development of each subfabric is dependent upon the response of the particles or crystals defining that subfabric within the deformation regime. Generally, in a non-coaxial strain regime, oblate markers with lower axial ratios have a higher angular velocity towards the shear plane than those with higher axial ratios, because the former have a smaller period of rotation (see above; Willis 1977; Fernandez & Laporte 1991). Sense of shear may be inferred from the angular relationship between different subfabrics within the XZ plane, where the subfabric defined by crystals with high axial ratios is visually rotated towards the subfabric defined by crystals with lower axial ratios (see Fig. 4.16), since this is how the fabrics would have continued to evolve if the

magma had not crystallised. Fernandez and Laporte (*op. cit.*) applied this technique to fabric analysis in pole figures, and proved that it is applicable on both the mesoscopic and microscopic scale.

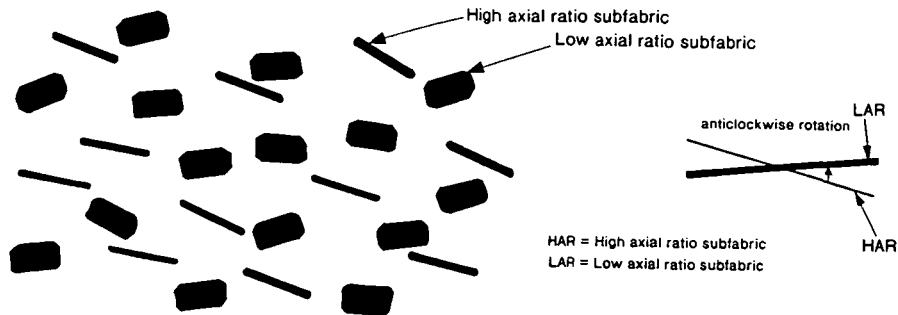


Figure 4.16 Sketch diagram to demonstrate the obliquity of rigid markers of different axial ratios deformed in the magmatic state (adapted after Fernandez & Laporte 1991). Sense of shear can be determined by visually rotating the high axial ratio markers towards those of low axial ratio.

In addition to determination of sense of shear, Fernandez and Laporte (*op. cit.*) observed that density maxima often developed around a particular orientation. They determined that the strength of the density maxima is related to the amount of shear strain (γ), up to a critical value (γ_c), and with decreasing angle between the foliation plane and the shear plane. The latter angle may, in turn, be related to the amount of shear strain. Once a foliation has passed through the shear plane, or the amount of shear strain exceeds γ_c the density maxima has been observed to decrease (Fernandez & Laporte *op. cit.*; Ildefonse *et al.* 1992). These authors have also noted that highly inequant markers tend to define density maxima more frequently than markers with more equant shapes.

There are certain assumptions that are inherent to the method outline by Fernandez and Laporte (1991):

- (1) there is no mutual interference between phenocrysts, this requires that,
- (2) the concentration of phenocrysts within the magma is low,

(3) the subfabrics are in their first cycle of evolution.

The latter of these three assumptions is very difficult, if not impossible to determine in the field, and this should be taken in to account when analysing the shear sense data collected using this technique. The first and second assumptions can, however, be confirmed from simple field observations.

4.5.2 Application of the technique to the Thorr Granite

The author applied the technique outlined by Fernandez and Laporte (*op. cit.*) in the Thorr Pluton, since the granite pavement exposed horizontal surfaces that are thought to be roughly parallel to the XZ plane of the finite strain ellipsoid. Subfabrics are defined by different families of feldspar phenocrysts; one subfabric is mainly defined by alkali feldspar and sometimes plagioclase feldspar phenocrysts that have low axial ratios, the second subfabric is mainly defined by alkali feldspar phenocrysts with higher axial ratios, and sometimes by clusters of biotite and pyroxene. As to the assumptions inherent in the method, the phenocryst concentration appears to have been relatively low during the formation of the magmatic state deformation fabrics, since tiling fabric, that would indicate mutual interference between phenocrysts is rarely seen.

The method in the field involved identifying the subfabrics and often marking them with different colours. The long and short axes of the phenocrysts defining each subfabric were then measured in order to calculate the axial ratios. The orientation of each crystal within a designated sample area ($\leq 1\text{m}^2$) was then measured using a compass clinometer, and an average was measured within each of the sample areas; these are marked on figure 4.17. A preliminary sense of shear was determined by visually rotating the feldspar subfabric defined by crystals with high axial ratios towards the subfabric defined by crystals with low axial ratios; this was noted for each of the sample areas. The results of analyses of 25 sample locations from around the pluton are included in the appendix to this thesis. These results were used to draw rose diagrams (see Fig 4.18a-x) of the subfabric orientations in each of the sample locations shown in figure 4.17.

The author also collected orientated specimens from selected localities throughout the pluton for thin sectioning. Some of these thin sections were then projected onto tracing paper, and maps of their

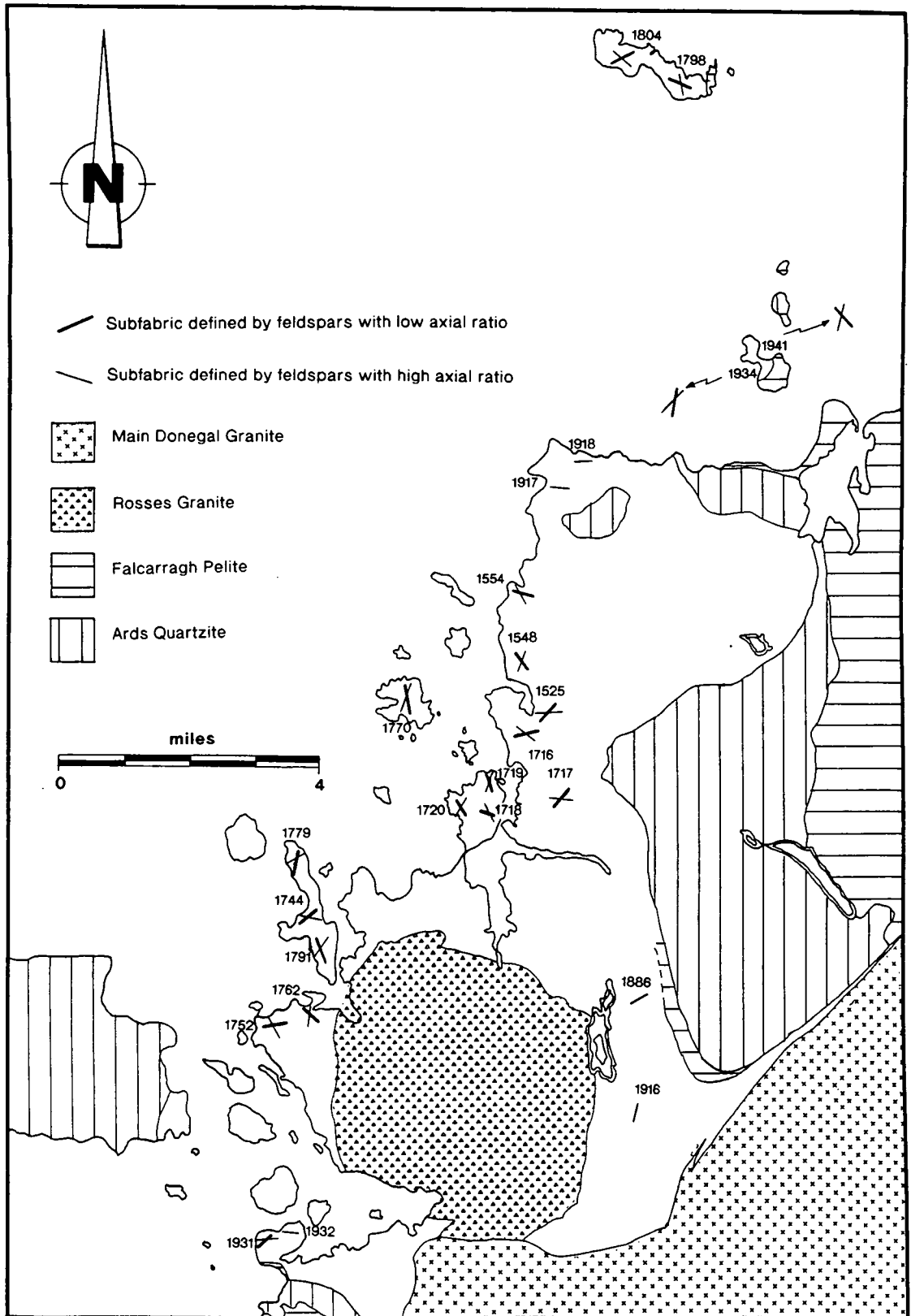


Figure 4.17 Sketch map of the Thorr Pluton to illustrate the general obliquity of magmatic state subfabrics at sample localities where detailed analyses were carried out.

subfabrics were drawn. The orientation of each crystal within the thin section map was then measured relative to the dominant subfabric maxima, as determined in the field. These results were also plotted as rose diagrams (see Fig. 4.19a-d). Only a few samples were studied in this way; this was to assess whether the technique is applicable microscopically without the use of crystallographic pole figures (see Fernandez & Laporte 1991 for application) and also to test whether the results found in thin section were consistent with those observed in the field. Unfortunately, the coarse grain size of the Thorr Granite limits the application of this technique in thin section, since it is often difficult to obtain enough phenocrysts, even within large section (5cm x 3cm), to make the test statistically viable.

4.5.3 *Shear sense within the Thorr Granite as determined by obliquity of subfabrics*

At each of the sample localities (Fig. 4.17 & Fig. 4.18a-x) the sense of shear determined by the application of this technique consistently indicates *sinistral* shear, since subfabric B, defined by crystals with high axial ratios, nearly *always* lie clockwise of those defined by crystals with low axial ratios (as summarised in Fig. 4.16). This is also the case for the four thin sections that were subjected to the method. The general obliquity between the magmatic state deformation subfabrics can also be seen by comparing the contoured stereonet of this fabric within each subarea. In each case subfabric B (high axial ratios) lies clockwise of subfabric A (low axial ratios).

Unimodal fabrics are seen in the Thorr Pluton when there is only one obvious population of feldspar phenocrysts that have approximately equal axial ratios. In such cases the alignment of mafic minerals may be slightly oblique to the feldspar alignment, but this occurs on too small a scale to be determined accurately with the naked eye, so no reliable shear sense can be determined in areas with dominantly unimodal magmatic state fabrics.

However, the orientation of the subfabrics does not remain fixed throughout the pluton, instead they seem to be arranged so that subfabric B, which forms the dominant subfabric with the strongest density maximum in many cases, is generally subparallel with the pluton margins (Fig. 4.17). This is particularly well-illustrated in the north of the pluton on Tory Island. In addition to this apparent parallelism with the pluton margins, the granite subfabrics also define a roughly concentric shape close to

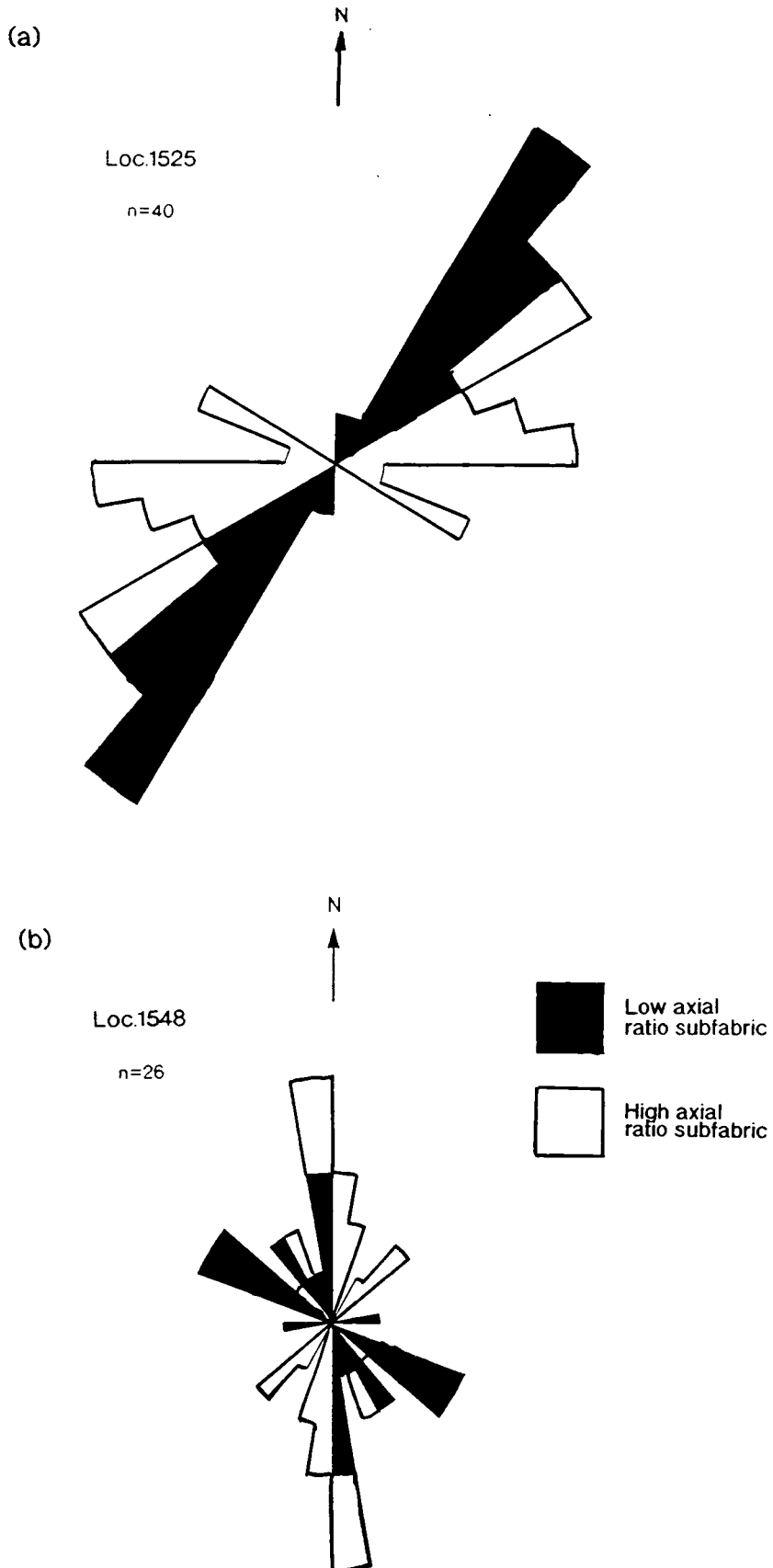
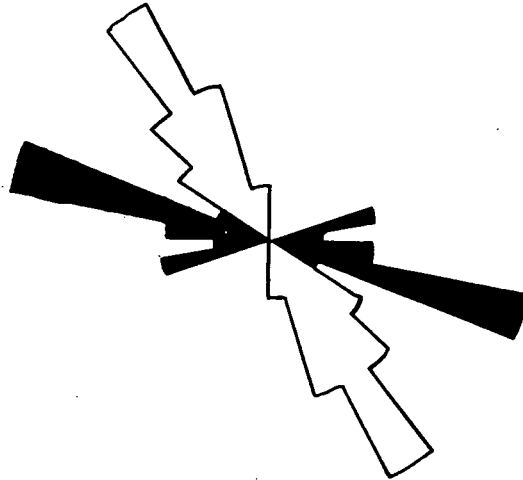


Figure 4.18 Rose diagrams of orientation analysis of phenocrysts of different axial ratios from selected localities around the Thorri Pluton. In most cases phenocrysts of low axial ratio lie anticlockwise of high axial ratio phenocrysts; this is consistent with magmatic state subfabric evolution synchronous with a component of sinistral shearing.

Figure 4.18

(c)

Loc.1554
n=26



(d)

Loc.1716
n=30

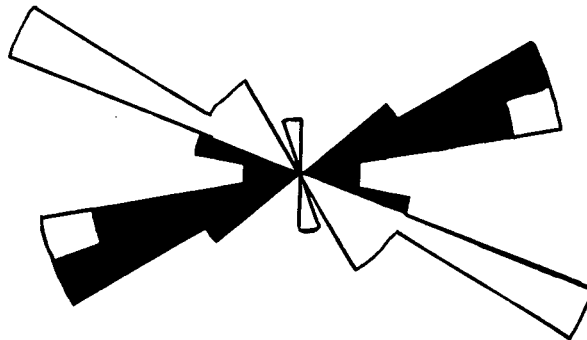


Figure 4.18

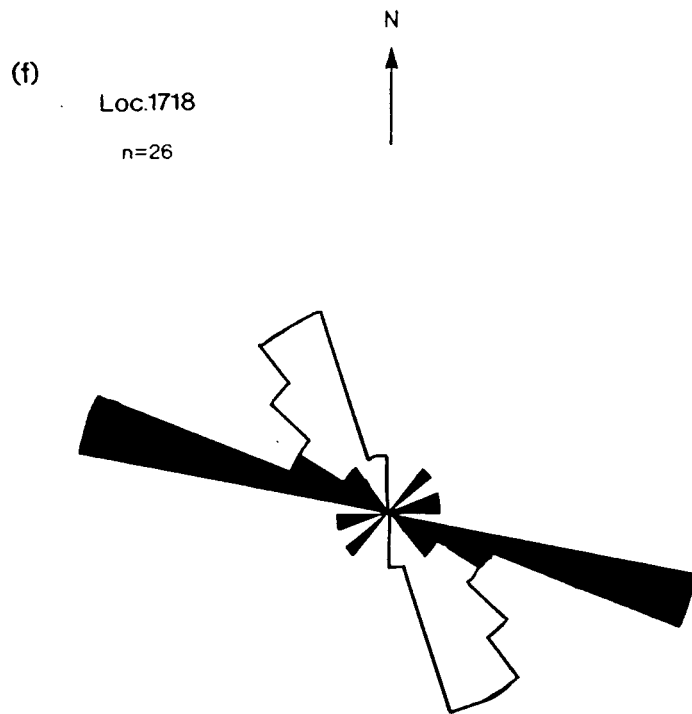
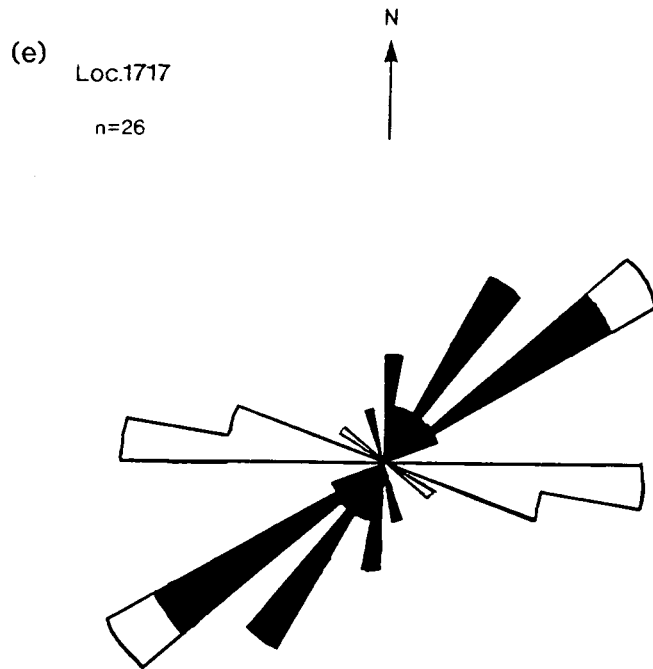
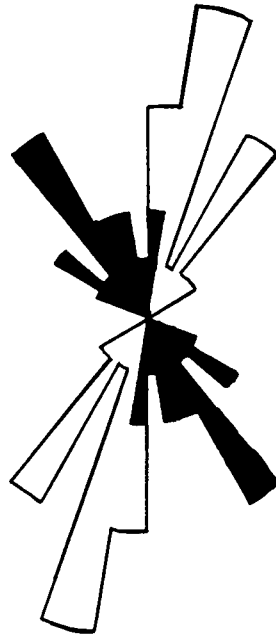


Figure 4.18

(g)

Loc.1719

n=24



(h)

Loc.1720

n=25

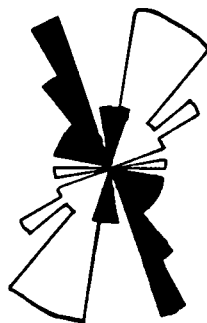
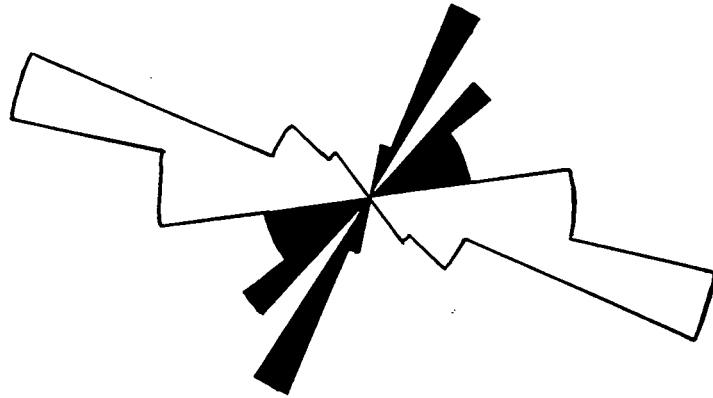


Figure 4.18

(i)

Loc.1744
n=34



(j)

Loc.1752
n=20

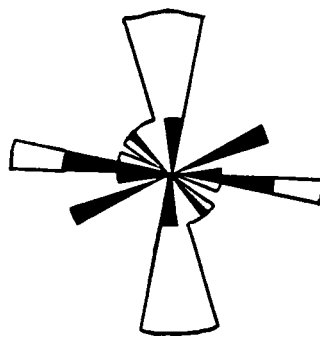
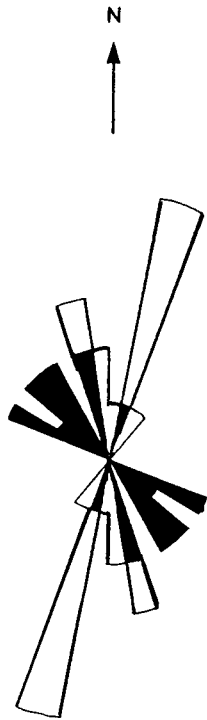


Figure 4.18

(k)

Loc.1762

n=20



(l)

Loc.1770

n=22

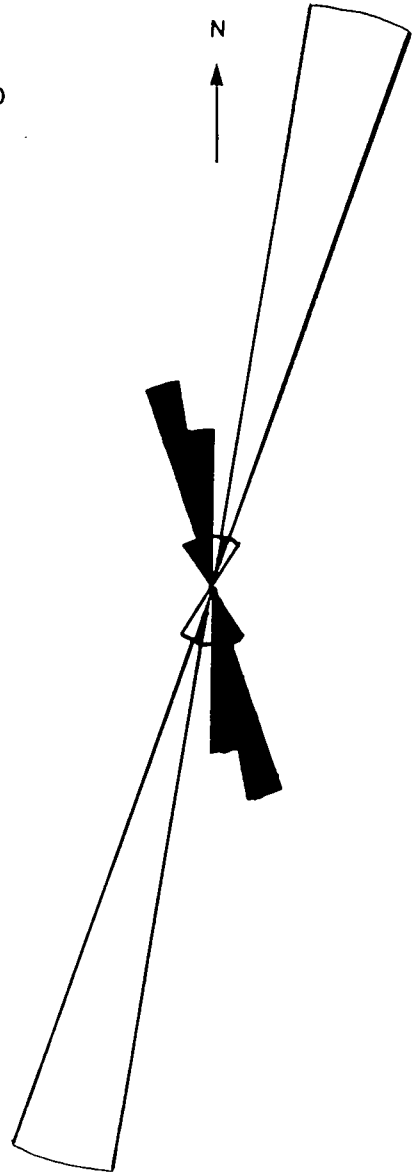
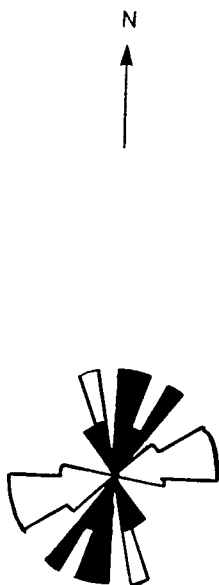
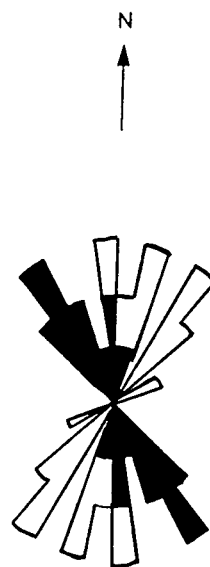


Figure 4.18

(m) Loc.1779
n=20



(n) Loc.1791
n=22



(o) Loc.1798
n=23

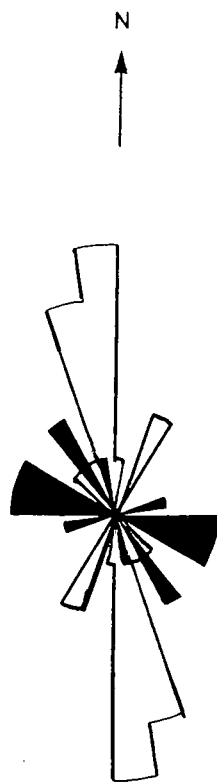


Figure 4.18

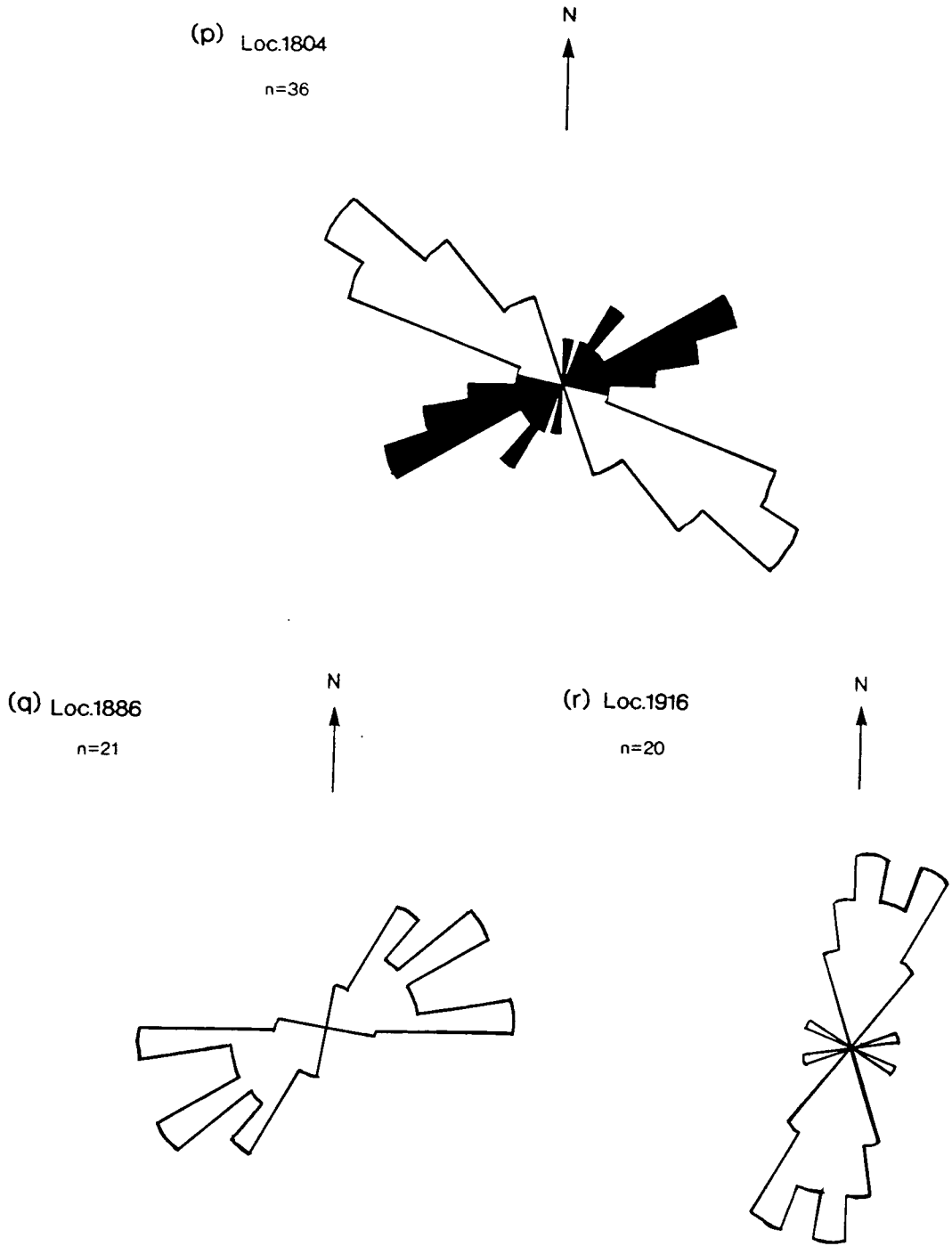


Figure 4.18

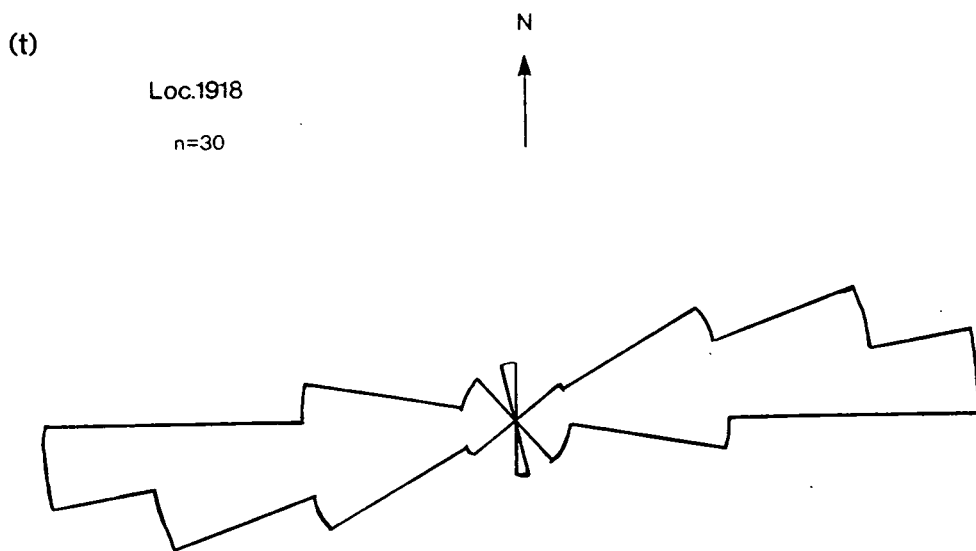
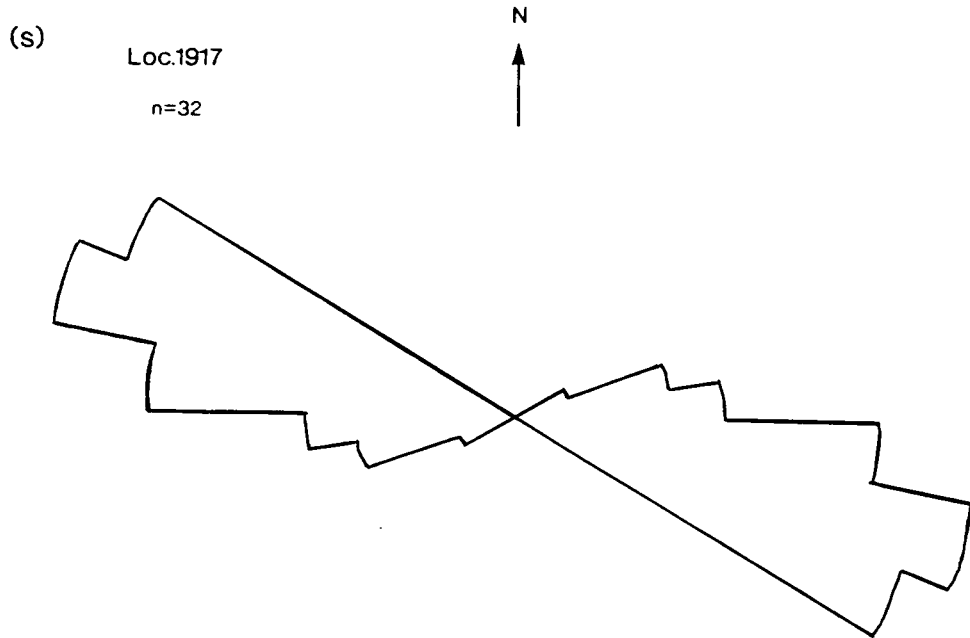


Figure 4.18

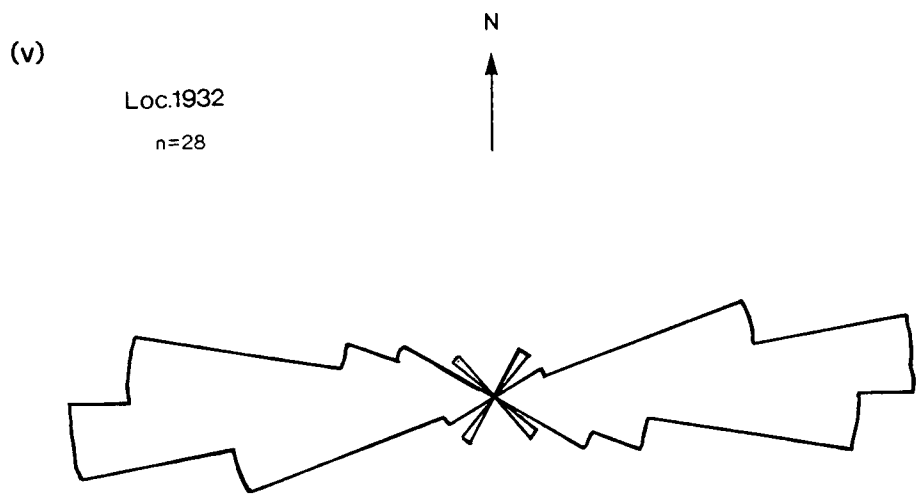
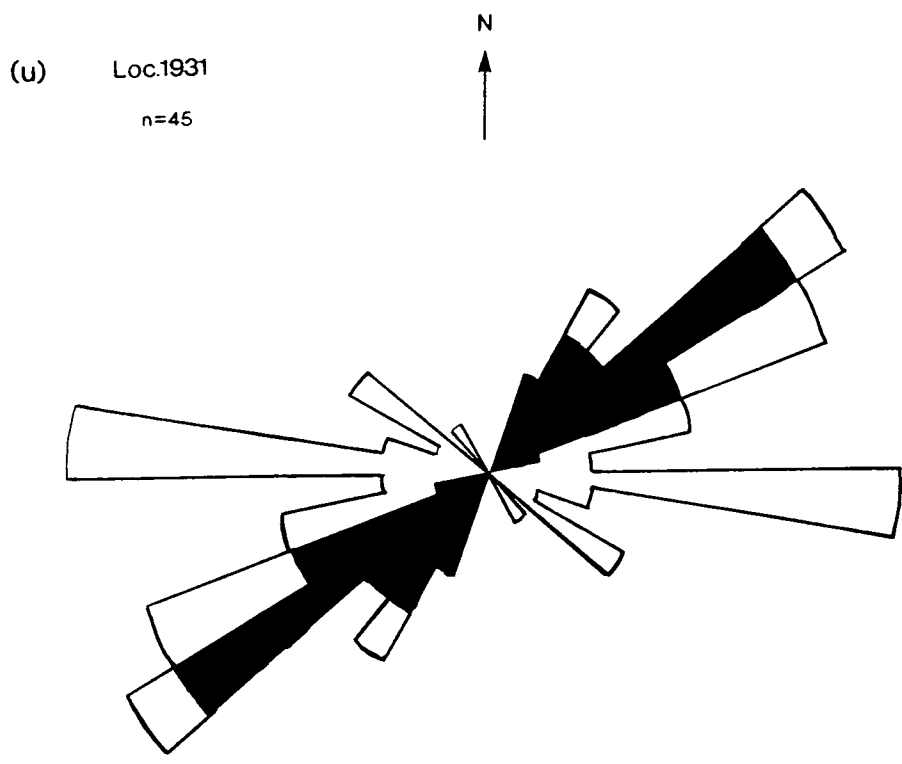
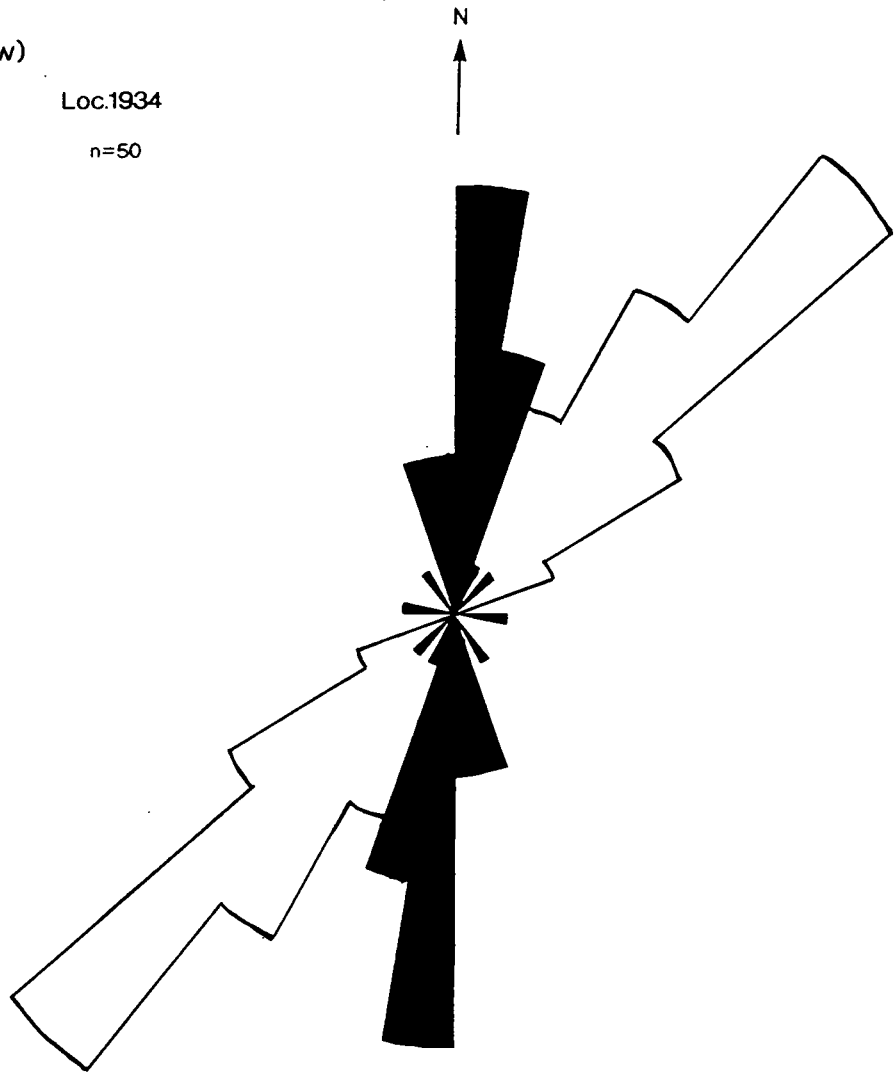


Figure 4.18

(w)

Loc.1934

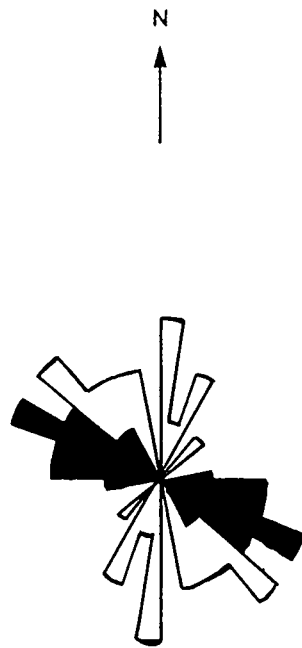
n=50



(x)

Loc.1941

n=26



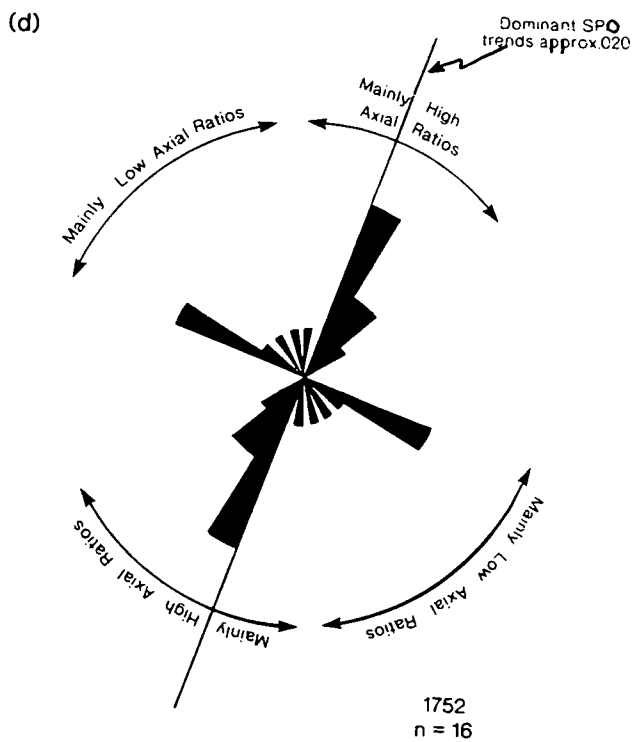
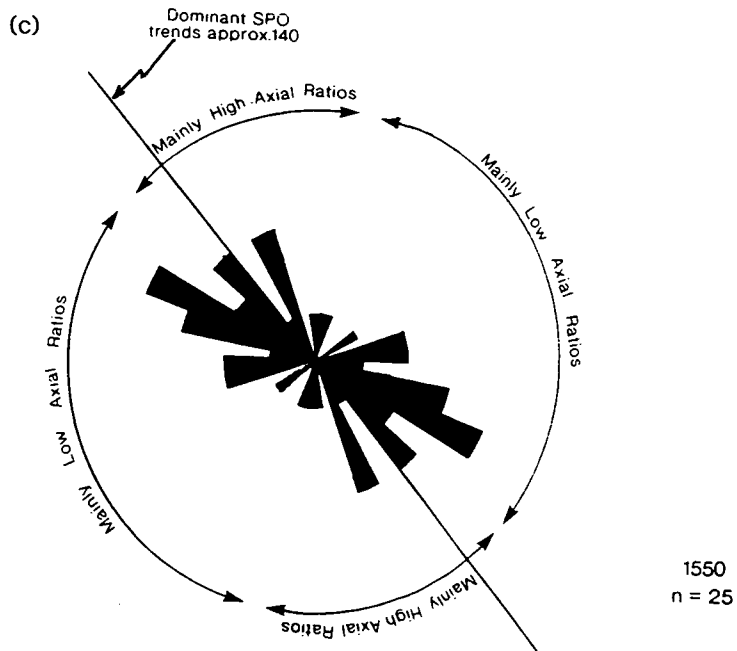
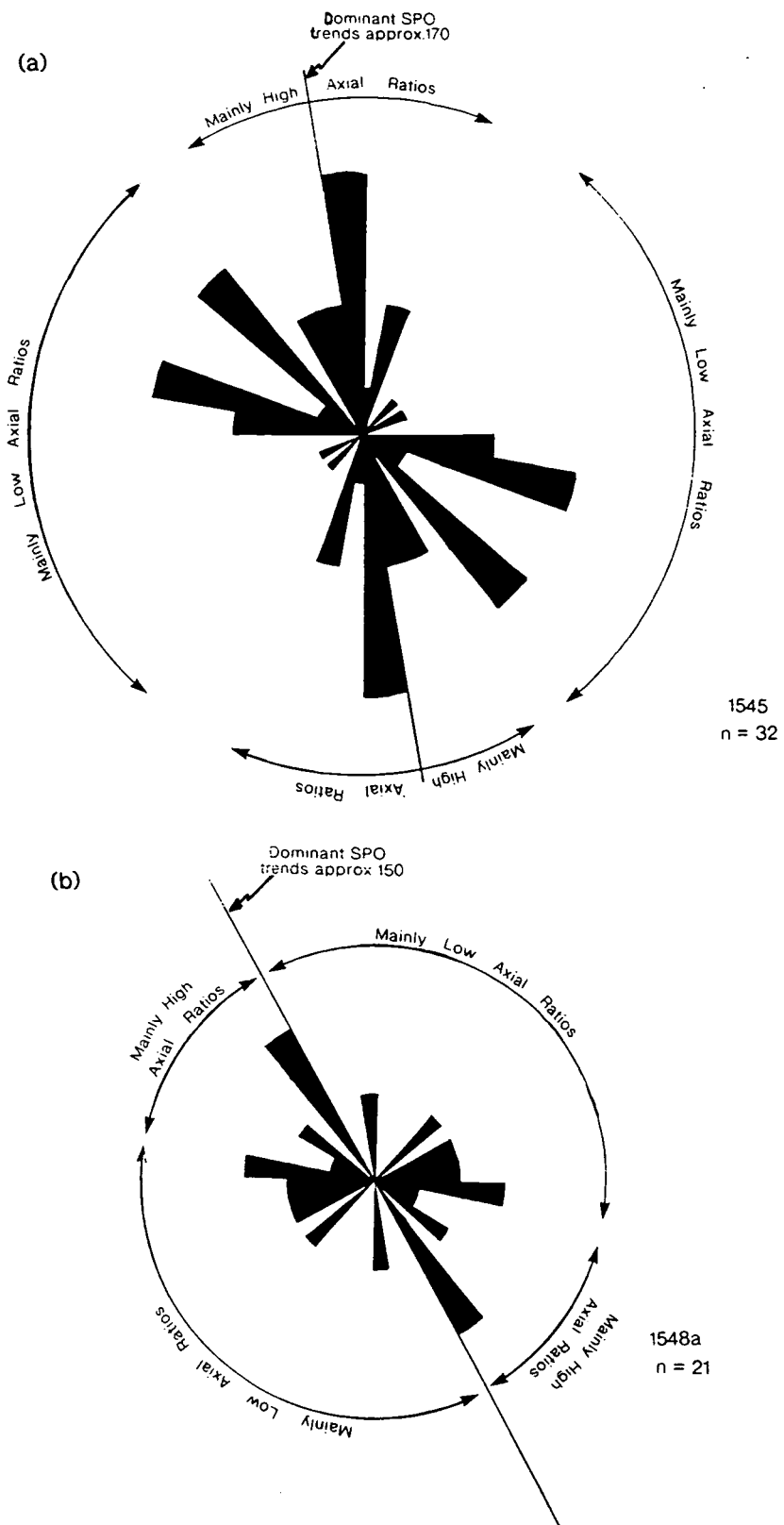


Figure 4.19 Rose diagrams of orientation analysis data for phenocrysts of different axial ratios from sample thin sections. In most cases phenocrysts of low axial ratio lie anticlockwise of high axial ratio phenocrysts; this is consistent with magmatic state subfabric evolution synchronous with a component of sinistral shearing.

Figure 4.19



the centre of the pluton (Fig. 4.17). This roughly coincides with the Gola Facies (Whitten 1957b; Pitcher & Berger 1972; Oglethorpe 1987) and the Inishmeane Centre (Oglethorpe 1987), both of which have been suggested as sites of emplacement of slightly later, felsic pulses of the Thorr Granite. The significance of the observed fabric patterns will be discussed later in this chapter.

4.6 Estimation of Strain associated with the deformation of the Thorr Pluton

4.6.1 Estimation from measurement of deformed markers

Analysis of the axial ratios of deformed markers are often used in order to determine the nature and the magnitude of the finite strain during a deformation event, in terms of the K value and R_{XZ} (Flinn 1965). This technique is the simplest of all strain analysis techniques (see Ramsay & Huber 1983 for review of most techniques) to apply in the field. It involves the measurement of the long and short axes of two dimensional sections through xenoliths in two or three orthogonal planes. The measurement planes must correspond as closely as possible to the principal strain planes (XY, YZ, XZ; Fig. 4.19). In order to ensure that the measurements and subsequent calculations are as statistically viable as possible, it is advisable to obtain at least 30 shape ratios which are then averaged to give a mean shape ratio (X/Y , Y/Z , X/Z) on that particular plane. Strains from any of the two principal planes can then be combined to determine the K value (see Ch.1). If, however, axial ratios can only be obtained from one of the principal strain planes, then it is impossible to determine the K value. In this case, only measurements from the XZ plane are of any value in analysing the strain, since the ratio of X/Z reflects the magnitude of the strain, R_{XZ} .

The precision of results obtained from application of this technique relies on the assumption that many deformed markers within igneous rocks were initially spherical. Hutton (1988a) stated that strain calculations of this type were most consistent when the deformed markers were cognate xenoliths, since they are assumed to be of a similar viscosity, and therefore have a similar rheology to their matrix. In addition, it was assumed that these are more likely to have initially spherical shapes than xenoliths of

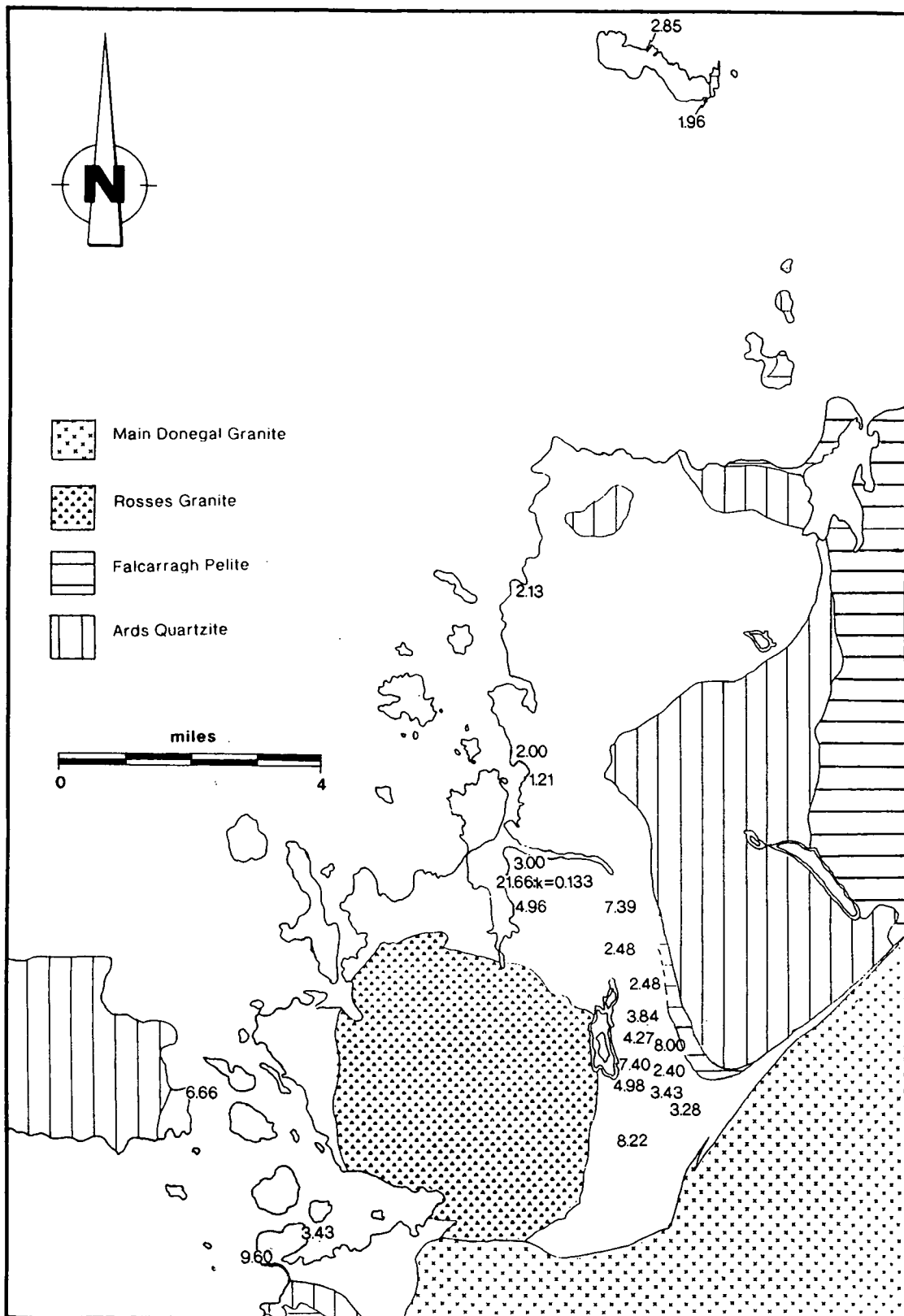


Figure 4.20 Sketch map of the Thorr Pluton with values of R_{∞} and K given at localities where elliptical markers were measured as strain markers.

country rock. The validity of the assumption that cognate xenoliths have initially spherical shapes will be discussed later in this chapter.

The Thorr Granite contains many xenoliths, many of which are xenoliths of country rock. Others, however, are magmatic in origin, and these are generally dioritic in composition. It has been suggested that these enclaves have originated from the appinitic magmas seen in the north of the pluton, and/or from synplutonic magmatic activity, as there is also evidence of this in the form of synplutonic dykes along the north and west coasts (Oglethorpe 1987). Unfortunately, these are often only measurable in one plane. Fortunately, this happens to coincide with the XZ plane, so at least the magnitude of strain can, in theory, be assessed. The measurements and the values of R_{xz} are listed in the appendix to this thesis. The values of R_{xz} are also summarised on the map of the pluton in figure 4.20. Typical values of R_{xz} vary from 1.2, up to 25, with the average being in the range of 2 - 8, which is the range of values found in the SE of the pluton, in the area around Thorr. This is the region where there is corroborative evidence of sinistral shearing synchronous with granite emplacement. Generally, values of R_{xz} tend to decrease to the N and W away from the Thorr area.

The largest values of R_{xz} , which are 16, 24 and 25 were recorded on the west coast, near the village of Rannafast (see Map 2). At this locality it was possible to measure the axial ratios of xenoliths in more than one of the principal strain planes, thus facilitating the calculation of a K value. The K value at this locality has been calculated as 0.133, which means that it lies well within the flattening field of the Flinn Plot (Flinn 1965). The fact that K is less than 1 implies that either the strain that resulted in the deformation of these markers was not a plane strain, or that the markers were not spherical to begin with, or some combination of these factors.

4.6.2 Estimation of shear strain around the pluton

Shear strain (γ) synchronous with granite emplacement in the area around Thorr (Maps 1& 2), assuming homogeneous simple shear, can be crudely estimated using the equation of Ramsay and Graham (1970):

$$d = \gamma w$$

this can be rewritten to:

$$\gamma = d/w$$

where d is equal to displacement along the shear zone and w is equal to the shear zone width.

Rough values of d and w were deduced based on offsets along sinistral shear fractures filled with granite along the SE marginal region of the pluton (see Ch. 3). These can be used in the above equation to calculate γ . The rough calculations are outlined below:

(A) for a 100m wide shear zone:

(I) Lower limit = 5m displacement,

$$\gamma = 5/100$$

$$= 0.05$$

(II) Upper limit = 50m displacement,

$$\gamma = 50/100$$

$$= 0.5$$

(B) for a 500m wide zone:

(I) Lower limit = 25m displacement,

$$\gamma = 25/500$$

$$= 0.05$$

(II) Upper limit = 250m displacement,

$$\gamma = 250/500$$

$$= 0.5$$

If we assume a simple shear model, the graph in figure 4.21 can then be used to determine the range of R_{xz} values and the predicted angle (θ') between the shear plane and the X axis of the finite strain ellipse for the calculated shear strains. The values of γ , θ' and R_{xz} are summarised in Table 4.1. The validity of the data outlined in Table 4.1 will be discussed in section 4.7.2 of this chapter.

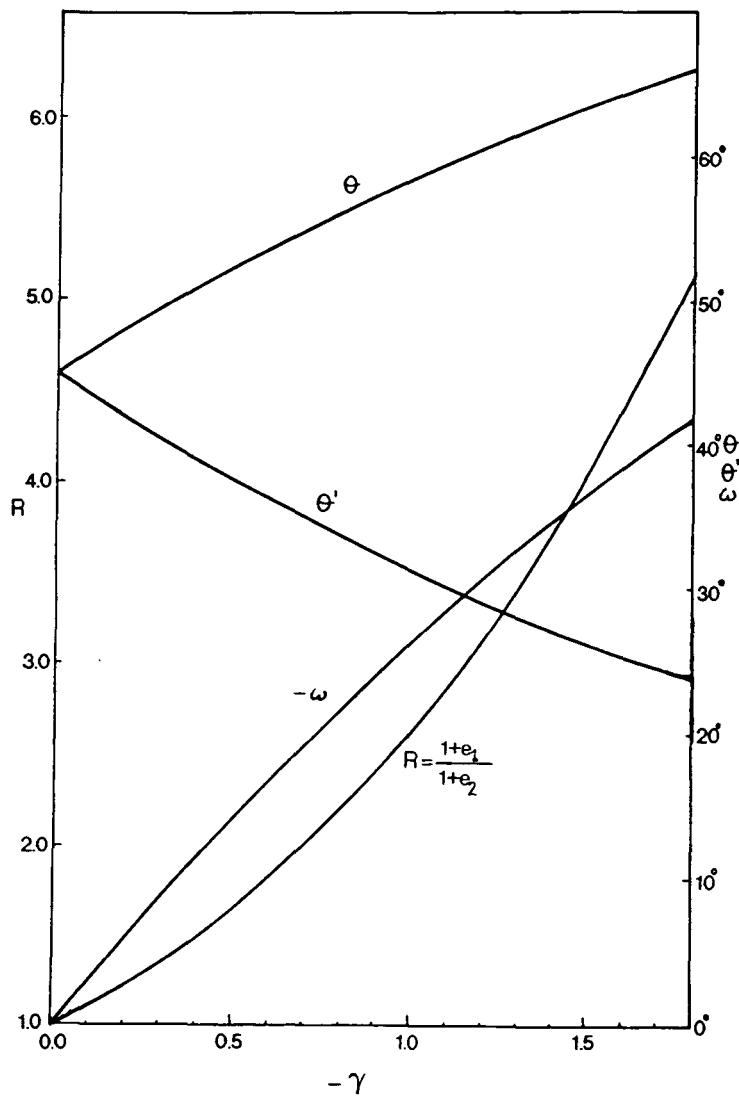


Figure 4.21 Variations in initial θ and final θ' orientations of the finite strain ellipse axes, ellipticity R of the strain ellipse and rotation ω as a result of progressive simple shear $[\gamma_{yx}]$ (after Ramsay & Huber 1983).

Table 4.1 Estimated values of γ , θ' and R_{xz} synchronous with the emplacement of the Thorr Granite.

	γ	θ'	R_{xz}
AI/BI	0.05	44.5	1.05
AII/BII	0.50	38	1.75

4.7 Discussion of granite textures, fabric geometry and evolution

4.7.1 Petrographical variants

Pitcher (1951, 1953a) and Whitten (1957b) proposed that the microcline microperthite and quartz in the Thorr Granite, in particular the Gola Facies, developed late in the crystallisation history of the pluton. They postulated that microcline microperthite was essentially a product of subsolidus metasomatic activity. They based this mainly on the fact that in thin section, crystals of alkali feldspar that appear to be euhedral in hand specimen are in fact subhedral, often filling interstices between minerals. The present author has confirmed the textural observations made by Pitcher and Whitten, but cannot support their hypothesis of formation, since the deformation textures within most of the granite, particularly the Gola Facies, are not consistent with deformation in the solid state (see section 4.3). The absence of solid state deformation fabrics in the Thorr Pluton generally, and more specifically in the feldspar minerals, which define the shape preferred orientation in the granite, suggests to the present author that these minerals must have crystallised from a liquid and have been deformed before this felsic portion of the Thorr Granite had reached its RCMP (see Ch. 1). This is in agreement with the work of Oglethorpe (1987) who believed that most of the minerals in the Thorr Granite, with the exception of epidote and secondary sphene, have textures consistent with having crystallised directly from a magma. The textural features of microcline microperthite can be explained by postulating that it started to crystallise *after* hornblende, plagioclase and biotite (see Ch. 2.6), all of which it includes, but before a substantial amount of the rock had crystallised, since the included minerals are small and the alkali feldspar occurs as phenocrysts. Microcline growth would have continued in a euhedral manner until it started to impinge on other crystals, at which point the alkali feldspar would start to grow around these other crystals, resulting in the microscopically subhedral and interstitial habit of the crystal exterior. This is consistent with the deformation textures within the granites. If microcline had grown by metasomatic replacement, one would expect that all deformation textures associated with it would be solid state features, but this is generally not the case. In the absence of such solid state textures, alignment of alkali feldspars that were produced by metasomatic replacement of plagioclase would have to be explained in

terms of mimetic overprinting of plagioclase after deformation had ceased. The author believes that the texture of alkali feldspar is consistent with crystallisation from a magma, and that the subhedral habit of the crystals can be explained if the late fractions of melt that filled the interstices was chemically similar to the alkali feldspar phenocrysts.

It is beyond the scope of this thesis to comment on whether the petrography the Gola Facies is indicative of it being a later pulse of more felsic magma, or simply the most evolved component of a magma that has undergone fractional crystallisation. However, the absence of any significant internal contacts in the area of the Gola Facies suggests to the author that the latter model is more likely.

The petrography of some of the granite specimens collected from the NE of the pluton, especially near to the contact at Binanea Strand, suggests that the granite here is somewhat more felsic and finer grained and equigranular than further south. Graphic intergrowths of quartz and feldspar are not uncommon, a texture which is interpreted as indicating emplacement at relatively shallow crustal levels.

4.7.2 Evolution and kinematics of the deformation fabrics and textures

4.7.2.1 Magmatic state fabrics

The bulk of the deformation fabrics in the Thorr Pluton are magmatic state fabrics, indicating deformation during, or shortly after emplacement. In general, more than one preferred orientation of phenocryst phases is developed, so that the bulk fabric elements can be subdivided into two, or sometimes more, distinct subfabrics which are oblique to one another. All subfabrics are steeply to subvertically inclined, but their strike varies considerably throughout the pluton (see below). Weak mineral lineations are developed throughout much of the pluton, most of which plunge shallowly within the foliation planes. The fabrics are typically $S > L$ tectonites (Flinn 1965).

The strike of magmatic state foliations appears to be strongly controlled by the orientation of the pluton margins. Close to granite - country rock contacts, the dominant subfabric, which is the subfabric defined by phenocrysts of higher axial ratios, becomes aligned subparallel with the contacts (see Map 2).

The weakest deformation fabrics are found in the NE of the pluton, close to Binanea Strand. Granite textures are here dominated by primary igneous textures; deformation fabrics are generally only

weakly developed. Some of the primary igneous textures, such as graphic intergrowth of quartz and feldspar and fine grain size, are consistent with granite emplacement at relatively shallow crustal levels and rapid cooling of the magma upon intrusion. The mode of emplacement of granite veins close to the contact at Binanea Strand is indicative of tensile hydraulic fracturing (Etheridge 1983), a process that could also be responsible for the formation of intrusion breccias. This suggests low values of differential stress which would also be consistent with relatively passive emplacement, and also with the lack of deformation fabrics in the granite. Furthermore, the calculated magnitude of strain (R_{xz}) associated with deformation during, or shortly after emplacement seems to be lowest in the north of the pluton (Fig. 4.20). However, accurate determination of the amount of strain using these data is open to question since the method requires that the deformed markers (cognate xenoliths) were originally spherical. However, at certain localities along the north and northwest coasts this was observed to be an invalid assumption, since enclaves frozen during separation from their host magmas had ellipsoidal, and even folded shapes (Plate 4.27). This implies that the calculated values of R_{xz} probably substantially *over-estimate* the magnitude of the deformation. It also calls into doubt the significance of the calculated K value of 0.133, indicating an oblate, or flattening strain, since the deformed markers measured in order to calculate this value may have been oblate to begin with.

However, if one accepts the strain data at face value, then calculated values of R_{xz} indicate that the deformation during, or shortly after granite emplacement was greatest in the S of the pluton, particularly in the area close to the sinistral shear zone observed in the country rocks. The magnitude of strain decreases away from this zone.

The crude estimates of shear strain (0.05 - 0.5, calculated in section 4.6.2) across the observed sinistral shear zone in the Thorr area are rather small and predict an angle between the X axis of bulk strain and the shear plane of between 44.5° and 38° . In reality, the angle between the magmatic state foliation and the shear plane in the Thorr area is generally less than the predicted values, but it seems to increase to the north and west of this area. However, in these areas the exact position of the shear plane is uncertain. Obviously the calculations used make many assumptions and are only crude estimates, but the data seem to suggest that syn-magmatic strains are highest in the region nearest to the sinistral shear zone



Plate 4.27 Folded dioritic enclaves within Thorr Granite exposed along the NW coast of Donegal.

seen along the main contact in the SE of the pluton at Thorr. It is notable, however, that there is no clear evidence preserved by granite fabrics in this SE region for the early phase of NW-side up shearing that produced initial forceful emplacement of granite, as recorded by the mobilised country rocks in this area.

Away from the Thorr area, the magmatic fabrics, particularly the subfabric defined by phenocrysts of high axial ratio, is subparallel with the contacts of the pluton. The absence of obvious shear zone in all but the extreme SE of the pluton suggests that there was little significant tectonic control of magma emplacement and fabric evolution in the north and west of the pluton. In such a situation, the author believes that the interaction between the pluton margin and the magma, with its associated fluid pressure, will ultimately control the orientation of the developing foliation. Such an interaction leads to the development of a 'buoyancy head' that is a product of the body force of the viscous magma acting against the rigid pluton walls. This is essentially a localised stress system that is capable of creating enough strain to produce a shape preferred orientation of early formed phenocrysts. However, if this is the case, and the magmatic fabrics observed in the north and west of the pluton are not directly attributable to tectonic deformation, then it is necessary to explain why the magmatic state subfabrics show a consistent sinistral sense of shear throughout the pluton. The author believes that this can be explained in terms of an inherited shear induced vorticity (Lister & Williams 1983) produced due to emplacement in association with sinistral shearing in the early stages of pluton emplacement, as preserved in the region around Thorr. Once a component of shear induced vorticity is imposed on a material, the material will continue to evolve with the same sense of vorticity until the momentum has been naturally dissipated, unless it is prevented from doing so, say by a superimposed and opposite sense of shear, or, more naturally, due to mutual interference of crystals in a cooling magma. Therefore, whilst the magmatic state fabrics in the Thorr pluton may not be exclusively the product of external tectonic forces, they do have a component of inherited shear induced vorticity that may originate due to the effects of tectonic forces.

4.7.2.2 *Solid state fabrics*

Solid state deformation fabrics are relatively restricted, but appear to be of three main types, each of different origin and relative timing. High to moderate temperature ($\geq 450^\circ$), coplanar solid state fabrics

are found in the contact facies rocks next to the main country rock - granite contact in the Thorr area and around the country rock rafts. In addition, localised patches of similar coplanar solid state deformation fabrics are sometimes found next to impersistent internal contacts within the main granite body.

The coplanar nature of these fabrics suggests that the deformation regime at the time of their formation was the same as when the magmatic state deformation fabrics were formed. This is consistent with a continuation of early deformation across the RCMP. The author believes that the solid state fabrics are localised in this way due to partitioning of deformation next to contacts between different granite facies and between granite and country rock. This may be due to the actual physical nature of the contacts, but it is more likely, particularly in the case of contacts between country rocks and granite, that elevated vapour pressures, particularly water vapour pressure, increased the potential for solid state deformation. This 'softening' effect will be especially well developed due to the relatively large proportion of quartz in the contact facies rocks. Water can lead to very significant hydrolytic weakening in quartz which would, therefore, lead to greatly enhanced solid state ductile deformation (Griggs 1967; Kirby 1984; Kerrich 1986).

Solid state deformation fabrics are seen in the Crovehy Hills (Map 2), where they are believed to be related to sinistral, transcurrent movement on the Main Donegal Granite Shear Zone (MDGSZ). Movement along this large shear zone initially resulted in anticlockwise deflection of the magmatic state fabrics within the Thorr Granite, resulting in their WNW orientation in the Crovehy Hills, as opposed to their more usual NNW orientation in the Thorr and Meencorwick areas immediately to the north. Deformation in this zone then produced a pervasive, sinistral S-C solid state overprint to the slightly earlier magmatic state fabrics up to 1.5km away from the MDGSZ. These solid state fabrics become localised into discrete and intensely deformed zones towards the main shear zone, ultimately resulting in steeply orientated, ENE trending brittle-ductile mylonite zones that carry an intense subhorizontal stretching lineation.

The present author believes that all the fabric and textural data from the Thorr Granite seen in the Crovehy Hills are consistent with a model in which the MDGSZ began to move before complete crystallisation of the Thorr Pluton. Deformation associated with the shear zone continued throughout the

crystallisation history of the pluton, producing fabrics and textures within increasingly localised zones in this retrograde system in the manner suggested by Gapais (1989).

Weakly developed, brittle-ductile solid state deformation fabrics are found inland of Maghera Bay, in the SW of the pluton (Map 2). In the field they appear to have a dextral sense of shear, determined from asymmetric wrapping of quartz. However, this could not be confirmed in thin section, so shear sense in this area remains somewhat ambiguous.

These are rather low grade fabrics, characterised by brittle deformation of feldspar and mainly, but not exclusively, ductile deformation of quartz, both of which act to reduce the overall grain size of the granite in this area. The apparently low temperature ($\leq 450^\circ$ for brittle deformation of feldspar; Simpson 1985) is consistent with evidence of low grade dextral shearing in the country rocks close to the granite contact north of Maghera Bay, indicating that there may have been a weak deformation event late in the cooling history of the Thorr Pluton. The exact origin of such an event cannot be commented on without further research, however, it is not unfeasible that it might be related to the emplacement of some of the later granite plutons in this area (e.g. Trawenagh Bay).

4.8 An emplacement model for the Thorr Granite

On the basis of the data outlined in this and the previous chapter, the following salient points can be noted about the Thorr Pluton.

(1) In the country rocks, the swing in strike from the usual NE-SW Caledonoid trend, to the NNW trend seen presently in NW Donegal had occurred *before* granite emplacement. In addition to anomalous strike, the rocks of the area around Thorr are very steeply dipping. The strike swing may have been a regional D_3 event, roughly coinciding with the production of F_3 folds, such as the Thorr anticline, the Crockator syncline and the Crockator anticline. The author could find no evidence to support the proposal of Rickard (1963) that the strike swing and some of the cross folds (see Ch. 2) were produced in response to forceful emplacement of the Thorr Granite.

(2) Many of the granite - country rock contacts preserve evidence of relatively passive emplacement of granite, involving *tensile* hydraulic fracturing under differential stress conditions of 4T or less. There is a strong tendency for intruding granite to utilise the dominant pre-existing anisotropy, for example bedding, crenulation cleavage planes and fold axial planes.

(3) In the SE of the pluton, the Thorr area, there is evidence of active tectonism in association with granite emplacement. Initial forceful emplacement is recorded by the production of the main phase folds in the pelitic country rocks that were mobilised due to increasing internal pore fluid pressures in response to dehydration reactions associated with thermal metamorphic effects of the granite. This forceful emplacement resulted in a *west side up* sense of shear. Pelites accommodated this strain by becoming folded, whilst quartzite accommodated the strain associated with initial forceful emplacement by top to the west shearing along the dislocation zone associated with the over-tightened Crockator anticline (Rickard 1963). Early folds within the country rock rafts strike slightly clockwise of those in the mobilised pelites due to clockwise rotation of the country rocks (towards the west) in response to early forceful emplacement of granite. Forceful emplacement was followed by *shear* hydraulic fracturing of the mobilised country rocks at differential stress conditions $\geq 8T$ and reactivation of the tectonic boundary between pelitic and quartzitic country rocks. The kinematics of fracturing and reactivation consistently indicate sinistral shearing.

(4) Most of the pluton carries a well developed magmatic state deformation fabric that becomes markedly bimodal in places due to the presence of phenocryst phases of different axial ratios that rotate at different rates in response to deformation. Coplanar solid state deformation fabrics produced by deformation in association with the emplacement of the Thorr Granite are localised within small zones in the contact facies of the granite around country rock rafts and the SE margin of the pluton. Other pervasive solid state deformation fabrics and localised mylonites are produced due to sinistral shearing along the ENE trending MDGSZ, in the S of the Thorr Pluton.

(5) The geometry of the magmatic state fabrics in the SE of the pluton is consistent with evolution within a sinistral shear regime. Away from this area, however, the magmatic state fabrics, particularly those with high axial ratios, become aligned roughly parallel to the margins of the pluton. Magmatic state

subfabrics have an obliquity that is consistent with anticlockwise rotation produced in response to sinistral shearing during fabric evolution. This is true even in areas where granite emplacement appears to have been relatively passive, and in such areas it is believed that the non-coaxiality of subfabrics can be explained in terms of a component of shear induced vorticity (Lister & Williams 1983).

(6) Deflection of magmatic state deformation fabrics, the pervasive solid state overprint in the S of the Thorr Pluton and localisation of solid state deformation into mylonite zones are consistent with initiation of sinistral shear along the MDGSZ *before* complete crystallisation and cooling of the Thorr Granite, with continued movement throughout the cooling history.

Given these salient points, the following model (sumarised in Fig. 4.22) is suggested for the Thorr Pluton.

Stage 1: Anomalous strikes and steep dips develop in NW Donegal in response to pre-granite intrusion deformation.

Stage 2: Hot magmatic fluids ahead of batches of magma get channelled along a precursor to the Main Donegal Granite Shear Zone (MDGSZ). These permeate along pre-existing tectonic contacts, such as the slide at Thorr that forms the contact between quartzite and pelite. This produces thermal metamorphic dehydration reactions in the pelites, resulting in an increase in their internal pore fluid pressure, and ultimately, to mobilisation.

Stage 3: Upwelling magma focuses into the 'softened' zone of mobilised pelites, where it is initially forcefully emplaced. The strain associated with early forceful emplacement results in the production of the main phase folds in the mobilised pelites. However, within the more competent quartzites the *west side up* sense of shear associated with forceful emplacement is accommodated along a dislocation zone in the hinge of the Crockator anticline. Early forceful emplacement and possibly some sinistral shearing along the precursor to the MDGSZ, also results in clockwise rotation of the country rocks towards the west.

Stage 4: ENE orientated sinistral shearing is initiated along the MDGSZ in response to the presence of magma at depth. This results in the development of a zone of transtension in the Thorr area, which when applied to the predominantly NNW trending strata produces a NNW trending cavity which magma can

the fill. The pre-existing tectonic contact between pelite and quartzite in this area, gets reactivated in response to transtension, and is the focus of localised sinistral shearing synchronous with granite emplacement.

Stage 5: Transtensile cavity continues to open, but the sinistral shearing remains localised in the Thorr area. Magma utilises the dominant pre-existing anisotropy to create more space for emplacement.

Deformation associated with emplacement results in the development of magmatic state fabrics.

Variations in the axial ratios of phenocryst phases that become aligned to define the magmatic state foliation, results in the development of subfabrics within the pluton. These subfabrics have a consistent sense of obliquity that suggests that during emplacement they inherited a component of anticlockwise shear induced vorticity in response to non-coaxial shearing. Early sinistral shearing along the MDGSZ result in deflection of magmatic state deformation fabrics within the Thorr Pluton.

Stage 6: Evolution of the pluton continues relatively passively towards the north. All magmatic state deformation granite subfabrics, including the relatively late Gola Facies, inherit a component of sinistral shear. Sinistral movements along the MDGSZ results in continued rotation of magmatic state deformation fabrics and country rock rafts up to 1.5km away from the shear zone.

Stage 7: As the pluton cools, coplanar solid state deformation fabrics develop locally in the Thorr area within the contact facies of the Thorr Granite. Pervasive solid state S-C fabrics and localised mylonite zones develop in the S of the pluton in response to continued movement on the MDGSZ throughout the crystallisation and cooling history of the Thorr Pluton.

Conclusions

- (1) The author cannot find support for the conclusion of Whitten (1957b) and Pitcher & Berger (1972) that alkali feldspar was produced by subsolidus replacement of plagioclase and quartz. The present study indicates that the textures of alkali feldspar and of the granite as a whole are consistent with crystallisation of alkali feldspar from a liquid *after* hornblende, plagioclase and biotite, which are included by alkali feldspar.
- (2) Thermal metamorphic effects resulted in dehydration reactions within the pelitic lithologies in the Thorr area, which increased the internal pore fluid pressures in these rocks and led to mobilisation.
- (3) The Thorr Pluton appears to have been emplaced initially forcefully; evidence for this is found in the country rocks, but not in the granite itself. Following this emplacement was associated with NNW orientated sinistral shearing occurred along a reactivated tectonic contact and within the mobilised pelitic metasediments in the Thorr area. Elsewhere around the pluton emplacement was rather passive and utilised the dominant pre-existing anisotropy. The tectonic regime can generally be described as *transtensional*.
- (4) Magmatic state fabrics were produced in response to deformation of the granite magma during, or soon after, emplacement. Subfabrics are developed where there is more than one population of phenocryst phases that have different axial ratios.
- (5) Strain associated with the formation of the magmatic fabrics was highest in the south of the pluton, coinciding with an observed zone of sinistral shearing within the mobilised country rocks, but decreases northwards. However, calculated strain data must be interpreted with caution, since the assumption that markers used in strain calculations were originally spherical is not always valid.
- (6) The geometry of the magmatic state fabrics within the pluton is generally concentric, with the subfabric defined by markers with high axial ratios following the pluton margins. This is believed to indicate that the body force or 'buoyancy head' created between the intruding magma and the relatively rigid surrounding country rocks *exceeds* the ambient tectonic forces in most of the pluton. The exception

to this is in the SE of the pluton, where magmatic state deformation fabrics lie clockwise of the inferred shear plane, in an orientation that is consistent with sinistral shearing during their formation.

(7) Granite subfabrics produced by differential rotation rates of phenocrysts with various axial ratios consistently indicate anticlockwise rotation senses. The author believes that this is due to inheritance of a component of shear induced vorticity due to sinistral shearing synchronous with granite emplacement.

(8) The method of Fernandez & Laporte (1991) to deduce sense of shear from rocks deformed in the magmatic state can be applied in the field. However, one must bear in mind the assumptions inherent to the model and account for these when interpreting the data. Like all methods to deduce shear sense, it should not be applied in isolation. Furthermore, results obtained from the application of this method give no direct indication as to the origin of the non-coaxial strains that have controlled the fabric development.

(8) Movement along the MDGSZ may have initiated very early in the history of the Thorr Pluton, playing an important role in the siting of granite emplacement. Sinistral shearing along it continued throughout the crystallisation and cooling history of the Thorr Pluton, resulting in rotation of magmatic state fabrics, production of solid state deformation fabrics that become locally mylonitic in the south of the pluton. The latter type of fabric evolution is consistent with continuous deformation in a retrograde (down temperature) system.

Part III

***COMPARATIVE STUDIES IN GRANITE EMPLACEMENT AND
DEFORMATION***

Chapter 5
THE RATAGAIN COMPLEX

Introduction

This author was involved in the latter stages of a NERC/BGS project involving the production of a memoir for the Kintail sheet (see summary version in Fig. 5.1). The primary phase of field mapping was carried out by other investigators, with Dr. D.H.W. Hutton examining the pluton in detail. The present author spent only a limited period of time in the field area studying the igneous complex on a reconnaissance basis, and was primarily responsible for the examination of deformation microstructures in thin section.

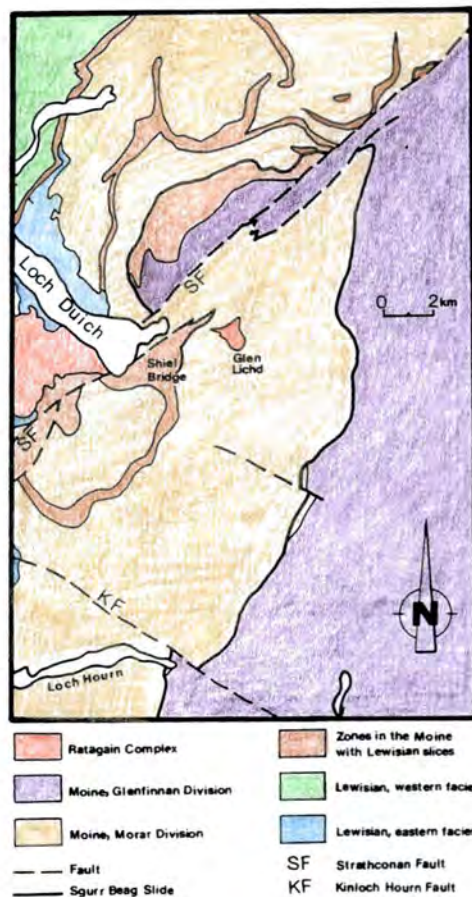


Figure 5.1 Summary version of the geology of the Kintail area.

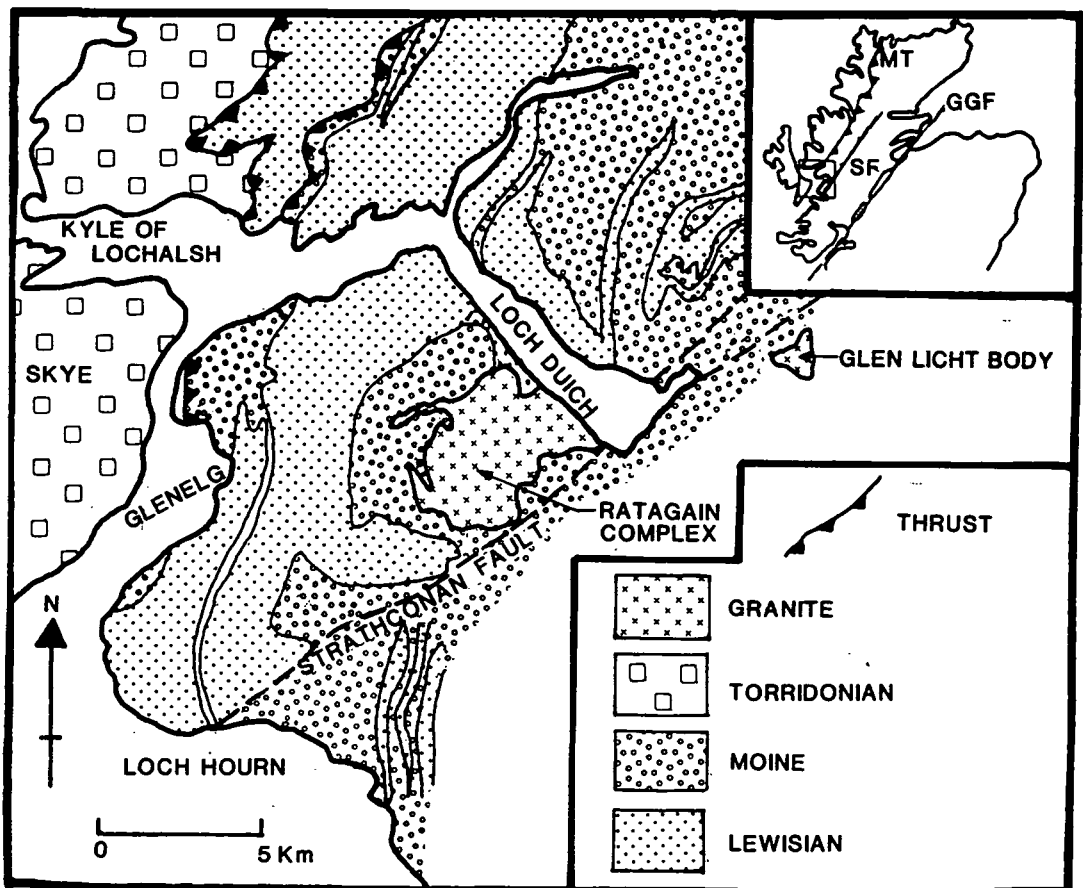


Figure 5.2 Location map showing the position of the Ratagain Complex and the general geology of the surrounding area.

The Ratagain granite complex is a small (17km²), deformed Caledonian pluton, in Inverness-shire, NW Scotland. It is emplaced into Moine and Lewisian rocks, adjacent to the Strathconan Fault, a member of the Great Glen Fault system, between the Sound of Sleat and Loch Duich (Fig.5.2). The Complex consists of three main plutonic units; diorite in the southwest, quartz monzonite in the northeast and small appinite bodies within the diorite and quartz monzonite units, the largest of these forming the south of the pluton (Fig.5.3). Minor variants on the main plutonic units range from melanocratic rocks associated with the appinites through syenites to true granites. Recent isotopic dating work by Rogers & Dunning (1989) has yielded a high-precision concordant U-Pb baddeleyite (ZrO₂) age of 425 ± 3 Ma for the pluton from a pyroxene mica diorite specimen. The granite complex and its country rocks were intruded at a slightly later date by a set of NW-trending lamprophyre dykes. These were dated by Rock (1988), using Rb-Sr techniques, at between 410-395 Ma.

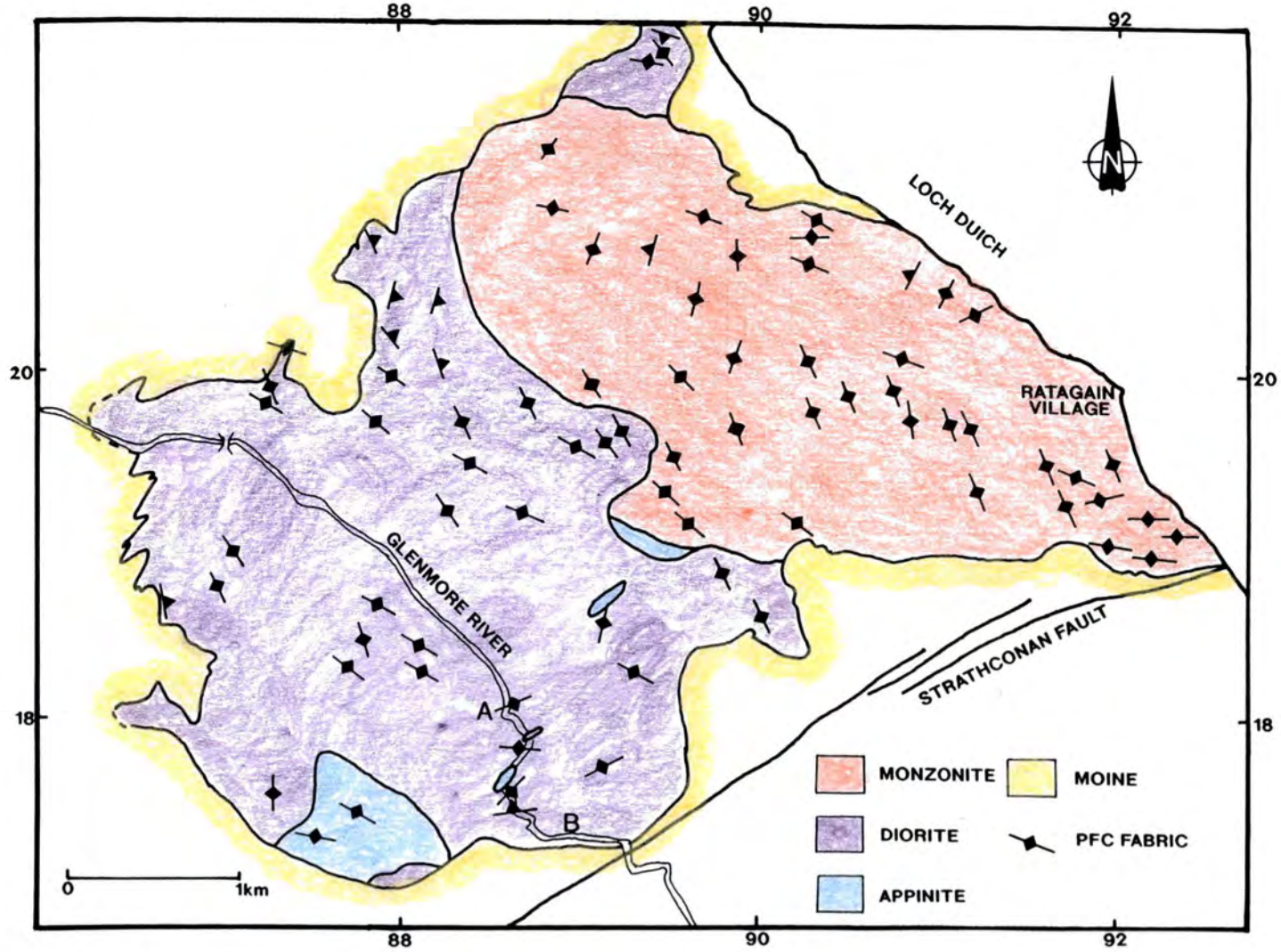


Figure 5.3 Major intrusive units and orientation of the magmatic state deformation fabrics within them. Also shows the position of the Strathconan Fault as exposed presently.

The well-constrained nature of the timing of intrusive events makes this pluton an ideal subject for studying the relationship between granite emplacement and regional deformation events, because it is also closely associated with important structural events in NW Scotland, including the waning stages of movement on the Moine Thrust and the development of regional sinistral shear zones associated with movements on the Great Glen Fault System. Furthermore, the pluton is important in a more general context of Caledonian magmatism, in that it represents a link or transition between the calcalkaline characteristics of the vast majority of the "Newer Granites" of Scotland and the alkaline characteristics of contemporaneous magmatism in the NW Highlands, including the syenitic intrusions of Assynt and Loch Loyal, the latter of which is the subject of Chapter 6.

5.1 Regional setting and structure

The igneous complex was emplaced into Moine metasediments tectonically interleaved with minor Lewisian units, the Glenelg inliers. It lies in the hanging wall of the Moine Thrust, the trace of which is located some 4.5km west of the pluton (Fig 5.4), whilst the Strathconan Fault lies immediately to the southeast of the pluton (Fig. 5.5). It is a significant member of the Great Glen Fault system with a protracted history of movement that will be commented on later in this chapter.

The Moine rocks of the northern Scottish Highlands form part of a deformed and metamorphosed Middle to Upper Proterozoic sedimentary sequence, which is tectonically bounded to the west by the Moine Thrust and to the southeast by the Great Glen Fault (Fig. 5.4, 5.5 & 6.2). The succession is dominated by psammites, semi-pelites and pelites that were laid down unconformably upon continental basement gneisses, which are thought to belong to the late Archean to early Proterozoic Lewisian complex (Holdsworth 1989 and references therein). The Moine and Lewisian were tectonically interleaved at various times and with differing styles, resulting in the occurrence of the Lewisian as inliers within the Moine outcrop (Peach *et al.* 1907; Johnstone *et al.* 1969; Rathbone & Harris 1979).

The Moine rocks are well known for their complex, protracted history of ductile deformation (e.g. Ramsay 1967). However, it was not until the 1980s that a coherent regional framework was

established due to the recognition and analysis of three related regional features of the geology; *reworking*, *basement-cover relationships* and *Caledonian ductile thrust tectonics* (see Holdsworth 1989 and references therein). Within this framework, structural and stratigraphic correlations can be constrained.

Reworking: Deformation and/or metamorphism of much, if not all of the northern Highland Moine during a Precambrian orogenic episode (*c.* 1000Ma) called the 'Ardgourian' (Barr *et al.* 1986) is strongly suggested by the results of isotopic dating studies (Powell & Phillips 1985). Subsequent to this early deformation, ductile deformation and associated metamorphism related to Caledonian orogenesis (*c.* 470 - 430 Ma) produced heterogeneous reworking of a large part of the 'Ardgourian' orogen. Much of the work over the last decade has attempted to use isotopic, structural and metamorphic data to distinguish Precambrian and Caledonian phenomena (Barr *et al.* 1986).

Basement - Cover relationships: Tectonic strain along the margins of Lewisian inliers (Fig. 6.2) has generally obliterated all traces of an unconformity with the adjacent Moine cover sequence. However, an unconformable relationship can be inferred from the presence of Moine basal conglomerates at Loch Carron, Glenelg, Strath Carnaig and Strathan Bay (Fig. 6.2; Peach *et al.* 1910; Ramsay 1958; Strachan & Holdsworth 1988; Holdsworth 1987, 1989). In areas of low strain, the Lewisian rocks often preserve definite evidence of igneous events and/or high grade metamorphism which occurred before the Moine sediments were deposited (Barber & May 1976; Moorhouse 1977; Moorhouse & Moorhouse 1977; Rathbone & Harris 1979). The Lewisian inliers occur in a range of tectonic settings and have been emplaced at various times. In the Morar Division rocks at Glenelg and Morar (Fig. 6.2), the basement units lie within the cores of large isoclinal folds of probable Precambrian age (see below; Ramsay 1958; Powell 1974). In contrast, other inliers, such as those at Scardroy (Fig. 6.2) are at a higher structural level in the sequence, forming ductile thrust slices along the Sgurr Beag Thrust, a major Caledonian discontinuity (Fig. 6.2; Tanner *et al.* 1970).

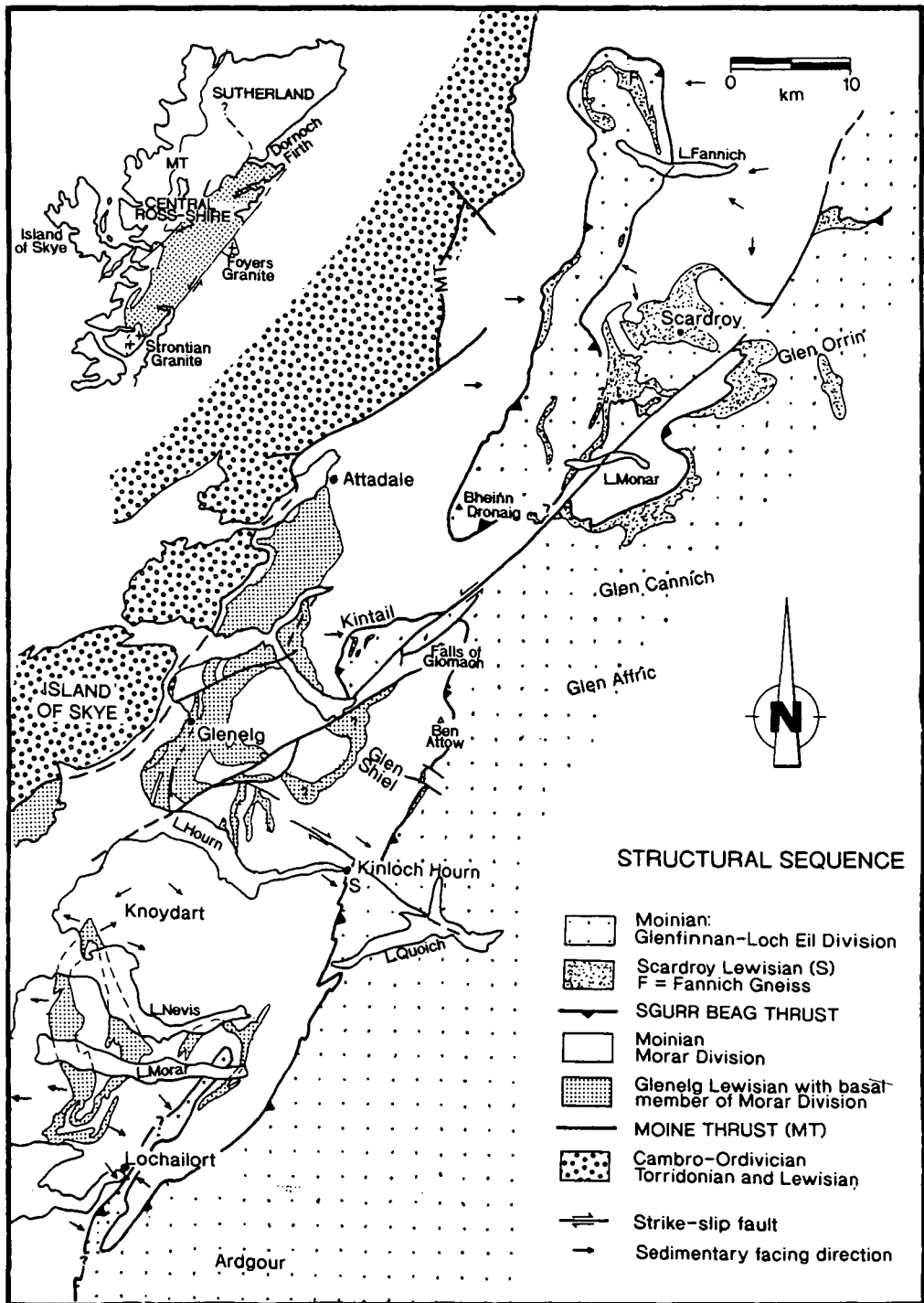


Figure 5.4 Stratigraphic and structural sequences for Central Ross-shire and part of Inverness-shire (after Tanner *et al* 1970).

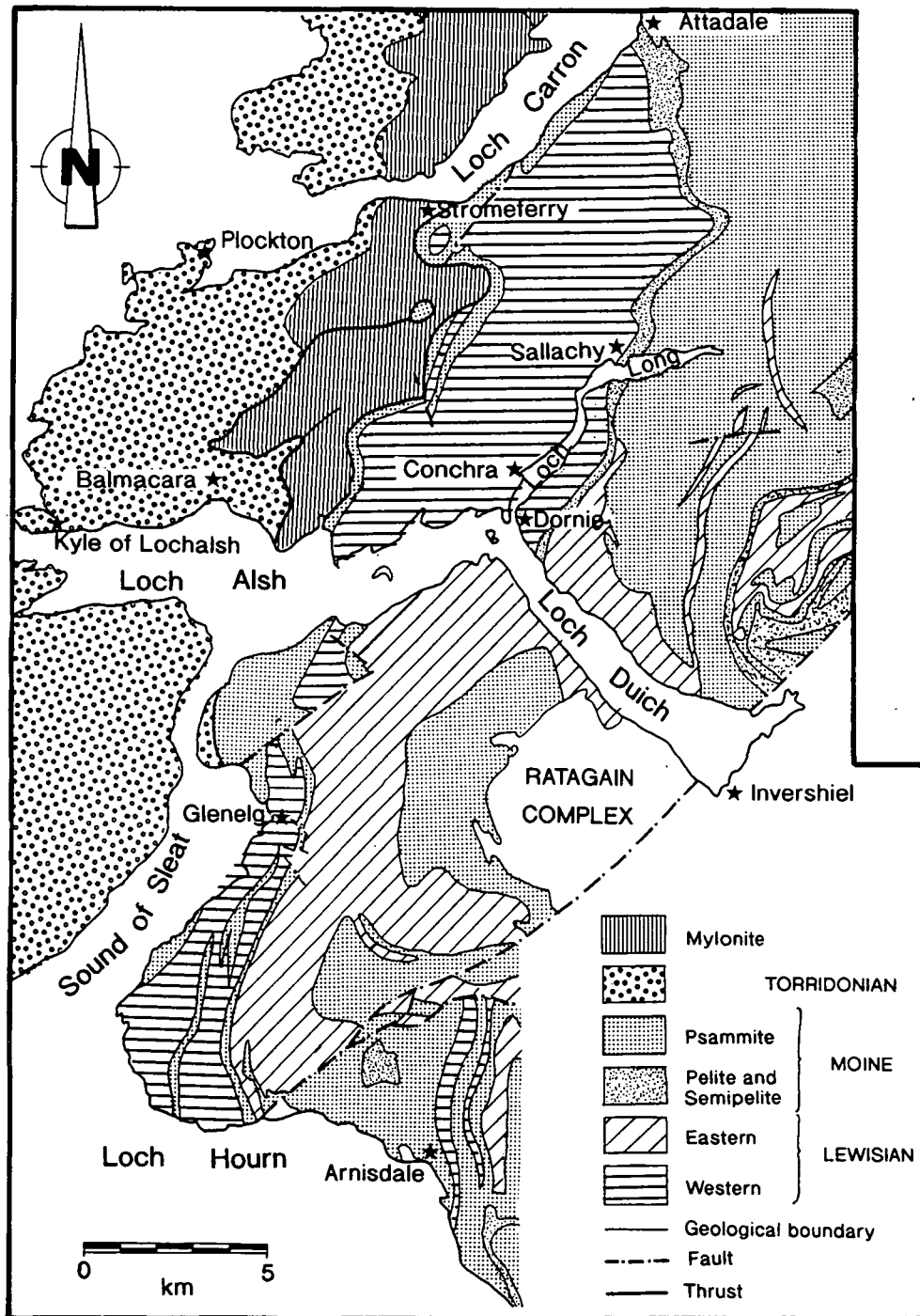


Figure 5.5 Detailed stratigraphic and structural sequences for Ross-shire and Inverness-shire (after Barber & May 1975).

Caledonian ductile thrust tectonics: It is now widely accepted that the northern Highland Moine forms an internal part of the Caledonian thrust belt, with the adjoining Moine Thrust Zone forming the external portion. A major breakthrough in the understanding of this complex area came with the recognition that the 'slides' within the Moine represent major mid-crustal ductile thrusts which are geometrically and kinematically analogous to the upper crustal discontinuities within the Moine Thrust Zone (Rathbone *et al.* 1983). As a result, Barr *et al.* (1986) proposed that the Moine rocks are sub-divided into three regional nappe units by the ductile thrusts; these are (from W to E) the Moine, Naver/Knoydart and Swordly/Sgurr Beag nappes (Fig. 6.2). Each of the tectonic units has unique structural, stratigraphic and metamorphic characteristics which collectively represent telescoped segments of the originally much larger Precambrian orogen.

5.1.1 Country rock stratigraphy and structure in the Glenelg area

The stratigraphy and structure of the Moine and Lewisian rocks of the Glenelg area have been the subject of much investigation during this century. The Glenelg inliers belong to a group of Lewisian bodies that occur as far south as Loch Morar and which are thought to lie at low structural levels within the Moine Nappe. They are considered to be separate from the overlying Central Ross-shire inliers (Tanner *et al.* 1970) that lie at higher structural levels in the hanging wall of the Sgurr Beag Thrust (Fig. 5.4). The interbanded Moine and Lewisian rocks are orientated parallel with the N-S trending inlier (Fig. 5.5, after Barber & May 1975). Along the western margin of the Glenelg inliers Lewisian rocks rest on a thin strip of Moine psammite which separates the inlier from the outcrop of the underlying Moine Thrust (Fig. 5.5; Barber & May 1975). Within the inlier, Lewisian rocks are separated into two areas, the eastern and western Lewisian, by a zone of highly deformed rocks, the central Moine strip of Sutton & Watson (1958). Barber (1968) proposed that this zone of high strain, which locally resulted in the development of mylonites, with a very well-developed ESE-plunging mineral lineation, represents the core of an isoclinal synform, which is also a tectonic boundary separating the highly deformed eastern Lewisian from the less deformed western Lewisian. The igneous complex was emplaced to the east of this tectonic boundary.

Clough (1910) recognised a basal conglomerate at the contact between the western Moine and western Lewisian belts southwest of Glenelg and showed that the Moine sediments were unconformable on the Lewisian. He also stated that both groups had

"passed together through a period of sharp isoclinal folding and intense metamorphism"

(1910, p.6)

and this folding was, in Clough's view, responsible for the alternation of the Moine and Lewisian belts, the Moine rocks occupying synclines between anticlines of underlying Lewisian. Clough attributed "later crumpling" to movements along the Strathconan Fault.

Kennedy (1955) adopted a modified version of Clough's interpretation by suggesting that the rocks of the inlier showed an imbricate structure and were over-ridden by a complex thrust sheet, the Morar Nappe, which outcrops south of Loch Houran.

Ramsay (1958), whilst agreeing with Clough's data, sought a different interpretation. He showed that the *observed* folding was not the cause of the interleaving of the Moine and Lewisian rocks. He ascertained that in the east of the area the Moine lies structurally below the Lewisian instead of occupying synclines above it. This, he believed was the result of very early interleaving of the Moine and Lewisian, which he initially attributed to thrusting. Sutton & Watson (1958) broadly agreed with Ramsay's proposal that interleaving of the Moine and Lewisian was a relatively early event and was succeeded by at least two later episodes of folding prior to movements on the Moine Thrust. However, later Ramsay (1960) and Ramsay & Spring (1962) changed the thrust model to one involving early isoclinal folding. The Moine stratigraphy was apparently symmetrical about these bodies and the highly deformed strip of Moine rocks between the Eastern and Western Lewisian could be seen to pass upwards into a synclinal structure in the less deformed areas. This is the traditionally accepted model and probably involved deformation which is pre-Caledonian in age since regional structural correlations in many areas demonstrate that the development of Caledonian structures such as the Sgurr Beag and Knoydart thrusts post-dates the development of these early, Lewisian-cored D_1 isoclines (Tanner 1971; Powell 1974; Powell *et al.* 1981; Baird 1982). Several authors (Brook *et al.* 1977; Brewer *et al.* 1979; Wilson & Shepherd 1979; Powell *et al.* 1981, 1983; Barr 1983; Kelly & Powell 1985) have used isotopic data to substantiate a pre-Caledonian

age for the early fold structures of Morar-Glenelg. However, some authors have gone further, in suggesting that the emplacement of all Lewisian inliers pre-dates the earliest recognisable structures throughout the NW Highlands (Mendum 1979; Smith 1979).

5.1.2 Fault Systems around the Pluton

(a) The Moine Thrust

The Ratagain complex is believed to have been emplaced at about 425 Ma during the later stages of movement on the Moine Thrust system (see Fig 5.4 & Halliday *et al.* 1987; Rogers & Dunning 1989, 1991). Watson (1984) proposed that the compressional movements on the Moine Thrust system had ceased before transcurrent movements on the Great Glen Fault system began. However, late extensional movements along the Moine Thrust Zone (Coward 1983), post-date the compressional movements, and may be penecontemporaneous with the initiation of strike-slip movements along the Great Glen Fault system, and cannot, therefore, at this stage be discounted as a possible factor in creating space for granite intrusion. Deformation textures within the plutonic rocks were examined (see section 5.4) to see whether they preserve any record of the final movements on the Moine thrust and whether it had any significant influence on the development of the igneous complex.

(b) The Great Glen Fault system and the Strathconan Fault

The Great Glen Fault is flanked by a 10km zone in which the regional strike is apparently deflected in a sinistral sense (Watson 1984). The Strathconan, Killin and Tay faults are associated with zones of unusual fold interference patterns. According to Watson (1984), these features suggest that the Great Glen Fault and its associated structures began to develop when the rocks which are currently exposed were still moderately ductile, with the first displacements being accommodated in broad shear zones, that narrowed into discrete fault zones as temperature decreased.

The Ratagain igneous complex is cut by the brittle Strathconan Fault (Fig. 5.5) and the Glen Licht body, which occurs on the SE side of the fault (c.f. Fig.5.1), may represent an offset part of the Ratagain Complex (Dhonau 1964). In this chapter, evidence will be produced for a protracted history of

movement along the Strathconan fault, not in a continuous manner, as suggested by Watson (1984), but rather as discrete events.

5.2. *The form of the pluton*

The Ratagain Complex has an exposed outcrop of approximately 17 km², with a presumed extension under Loch Duich based on aeromagnetic and gravity anomalies (Fig. 5.6; Hutton *et al.* 1993). Recent extensive road cutting associated with improvements to the Glenelg road over Mam Ratagain and forestry activities, have created substantial new outcrops of considerable value. Data from complete remapping of the complex has been combined with the pre-afforestation map of Clough (1910), by this author, to produce a new map. This exhibits significant differences from the published map (Nicholls 1951a) and the 1984 version of the 1:50 000 Sheet 72W. Differences relate mainly to the "Western Granite", the outcrop distribution of the appinites, the diorite - monzonite contact and the adopted petrological nomenclature. The petrological members of the complex were defined in this present study using the IUGS classification (Streckeisen 1976).

The outer contacts of the pluton are very poorly exposed, although they can be observed in several river sections, the Allt Sasaig, Allt na Muice and Glenmore river (Fig. 5.6 & Map 3). Here the contacts can be shown to be gently dipping, which is consistent with the outcrop pattern in the north, west and south of the pluton, where the contact follows the valley contours. In contrast, in the southeast of the pluton there is a suggestion that the contact is vertical or steeply inclined (see Fig. 5.6 & Map 3). Where the contact is steep, thermal metamorphic effects, visible to the naked eye, tend to be restricted to within 5m of the contact, with cordierite hornfels and evidence of local anatexis at contacts between diorite and Moine schist. There is, however, no evidence of any obvious change in the country rocks close to the quartz monzonite.

The "Western Granite" described by Nicholls (1951a) is not a mappable unit on the scale of Figure 5.6. It occurs as a very thin, discontinuous margin to the diorite in the west of the complex (Hutton *et al.* 1993). This has been interpreted on the basis of field, petrographical and geochemical

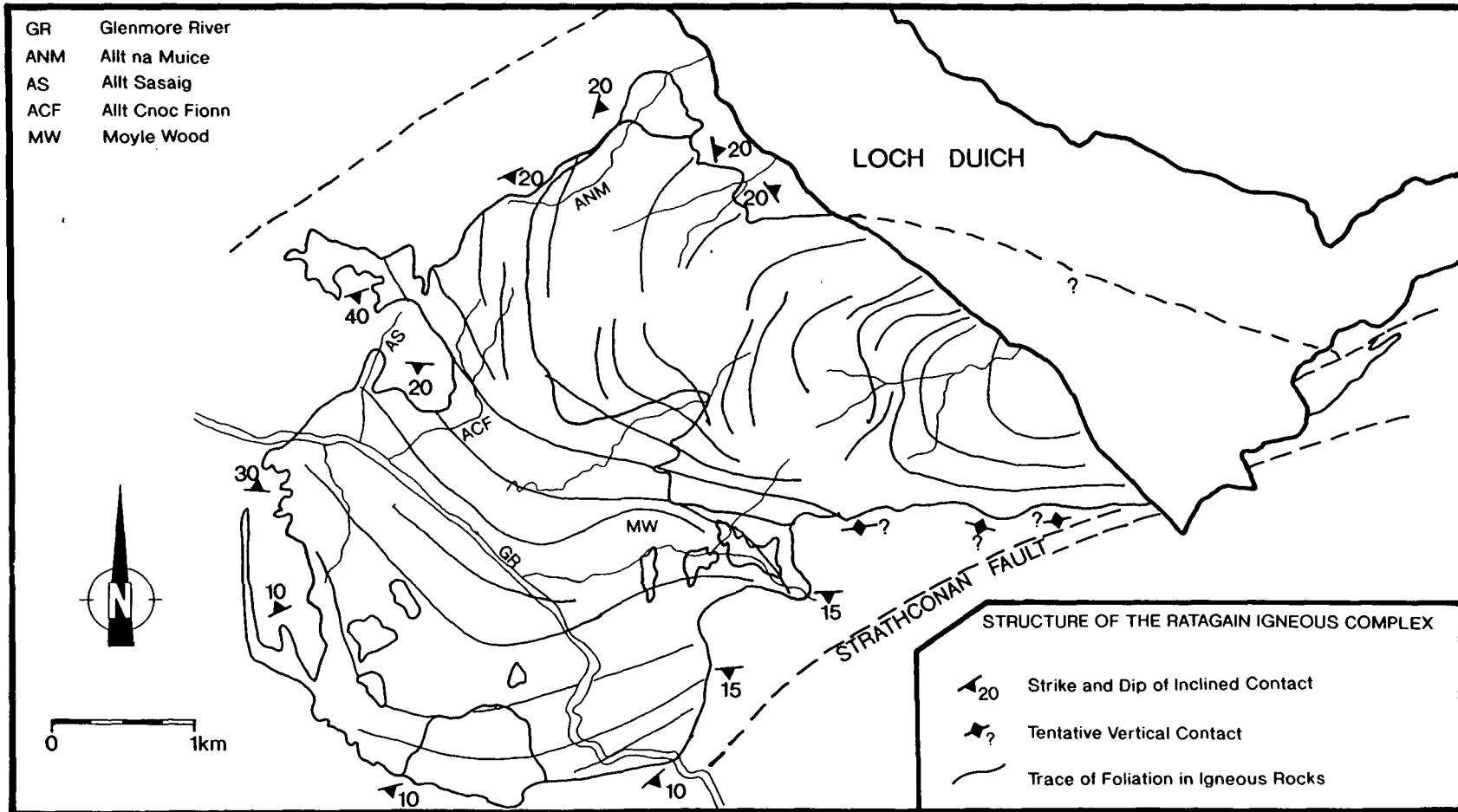


Figure 5.6 Sketch map of the Ratagain complex illustrating the dip of the contacts where exposed.

evidence to have been produced as a result of local anatexis in the Moine schists adjacent to the contact (Hutton *et al.* 1993). This interpretation is consistent with a shallow roof contact extending westwards, whilst the gravity anomalies on the Bouger Gravity Anomaly Map (1:250 000 Series 1976) and magnetic anomaly data on the Aeromagnetic Anomaly Map (1:250 000 Series 1978) suggest that beyond the roof zone the pluton contacts dip steeply, indicating that the overall form is one of a flat-topped pluton with steep sides (Hutton *et al.* 1993). The contacts are invariably sharp and cross-cut all folds and fabrics in the country rocks. Sheeting of granitoids into the country rocks is a very localised phenomenon, occurring only at the contacts.

5.3 The rocks of the igneous complex

The petrology of the complex was described briefly in the Geological Survey Memoir for Sheet 71 (Clough 1910) and a more comprehensive petrological investigation was published by Nicholls (1951a & b). These authors defined the principal units of the complex and their general petrological characteristics. The *c.*425 Ma emplacement age (Rogers & Dunning 1989), suggests that intrusion is broadly coeval with the Assynt alkaline suite and the Loch Loyal Syenite complex (van Breeman *et al.* 1979a; Halliday *et al.* 1987). A U-Pb age determination on a zircon from quartz monzonite by Aftalion *et al.* (1984) yielded an unrealistically young age of about 365 Ma which is now interpreted as reflecting some loss of radiogenic Pb from the zircons (Hutton *et al.* 1993). The Rb-Sr whole rock isochron age of 415 ± 5 Ma determined by Turnell (1985) was adjusted back to 419 ± 3 Ma by Thirwall (1988) and probably indicates fairly rapid cooling of the complex.

The main units of the complex are described in the following sections of text.

5.3.1 Diorite

The diorite is particularly variable and heterogeneous and has numerous small discoidal inclusions of amphibole-rich (appinite and meladiorite) material scattered through the body. In the field it is generally medium-grained, but variants range from fine- to coarse-grain size. Transitions between the

various grain sizes are usually gradational, although occasional sharp internal boundaries between different facies may be observed, for example in the Glenmore river (see Map 3 NC 886193).

Petrographically, the most common variety is a hornblende-biotite diorite, with accessory apatite, titanite and celestine. Locally, the diorite may contain pyroxene rather than amphibole and small outcrops of pyroxenite are exposed in the Glenmore river section. Olivine gabbro, which apparently represents the least evolved member of the complex (Hutton *et al* 1993), has been recognised along with pyroxene-mica diorite near the western outer contact in outcrops directly south of Braeside (Map 3).

Nicholls (1951a) indicated on his map that a large area of syenite separates the diorite from the quartz monzonites (the latter of which he called adamellites). In this area there are some diorites with a pink alkali feldspar-rich matrix, but modally these are better described as monzodiorites. Whilst there is evidence for the presence of syenite in this area, it is not a discrete mappable unit, but a distinct local facies within the diorite, with which it has gradational contacts. This syenite is often quartz-bearing (the nordmarkite of Nicholls 1951a) and typically contains clinopyroxene, amphibole (which may sometimes be of an alkaline variety), alkali feldspar and plagioclase, as well as quartz and accessory titanite. An example of this rock type outcrops on the north side of the Glen road, east of Allt Cnoc Fhionn (Map 3 NC 878197). Nicholls (1951a) noted the similarity between this nordmarkite, the perthosites of Loch Ailsh and the syenites of Loch Loyal, postulating that there must have been a supply of alkaline magma at depth for rocks of this type to be found in association with 'ordinary' calc-alkaline varieties. To facilitate the presence of peralkaline magma at depth, he speculated about the possibility of deep linked fault systems, which could act as a pathway for alkaline magmas from the Sutherland region further to the north, where they are clearly seen to be common.

5.3.2 Quartz Monzonite

This is essentially the unit that Nicholls (1951) described as adamellite. It occupies the core of the pluton and is normally a pinkish colour. In the west it is seen in contact with the slightly earlier diorite unit (diorite xenoliths are found within the quartz monzonite), whilst in the south the quartz monzonite is in contact with country rocks. The eastern margin is obscured by Loch Duich.

The contact with the diorite is well exposed in the recent Ratagain - Glenelg road cuttings. It is a zone some 200m wide in which the quartz monzonite intrudes and sheets the diorite in a steeply inclined contact zone. Within this zone the quartz monzonites normally contain both biotite and hornblende, but the overall mafic content is extremely variable, with a tendency to be higher towards the contact with the diorite. Close to this contact there is also abundant evidence of contamination by amphibole-rich material from discrete inclusions of appinitic and dioritic material down to the scale of single crystals.

Moving away from the contact, the composition grades into true granite in the centre of the quartz monzonite unit. The exact origin of this zoning, whether lateral or concentric petrological zoning, is not clear due to the fact that Loch Duich obscures the eastern margin.

5.3.3 Appinite Suite

These occur as both discrete pipe-shaped bodies and abundantly distributed enclaves in both the diorites and quartz monzonites throughout most of the outer portion of the pluton. In terms of the current IUGS nomenclature, the appinites of the Ratagain complex are meladiorites, melamonzonites and melasyenites. The mafic mineralogy is dominated by hornblende, biotite and occasionally pyroxene, with a total mafic modal abundance of up to 90%. Plagioclase to alkali feldspar ratio in the groundmass is very variable. Calcite, apatite and celestine are common accessories.

The best exposures of appinite outcrop in the Glenmore river and in the Forestry Commission road cuts in Moyle Wood. One such outcrop at NC 897 189 consists of pillow-like masses, 0.1 - 0.5m in diameter, of melamonzonite with their long axes aligned vertically in a matrix of pink monzonite, forming a vertical pipe-like structure (Plate 5.1). No pipes of appinite are observed to cut the centre of the monzonite unit.

The present study suggests that the appinitic magmas developed early and that a substantial body 'ponded' beneath the present level of exposure. This body was firstly cut by diorite and then by monzonite, both of which carry inclusions of the appinite. At a later stage the appinitic magma was itself intruded into the major plutonic units as a series of pipes, mobilised by the relatively volatile-rich quartz monzonite



Plate 5.1 Pipe-like appinites intruded into a matrix of pink monzonite as seen in Moyle Wood (NC 897 189).

magma. Nicholls (1951a), however, considered that at least one of the appinitic-like exposures in the Glenmore river was a skarn deposit. The current study found no evidence to support this.

5.3.4 *Minor intrusions*

Minor intrusions, generally in the form of dykes and steeply inclined sheets, are unusually abundant around the Ratagain complex (Smith 1979). Those occurring within the plutonic complex are predominantly members of the Microdiorite Suite, with sparse occurrences of intrusions of the Minette Suite (as defined by Smith 1979). Dykes of the Microdiorite Suite include hornblende-rich microdiorites, which are mainly found intruding the diorite, and very abundant felsites and porphyrites, which cut all members of the complex. The Suite appears to span the compositional range from diorite to quartz monzonite. The Minette Suite dykes constitute a swarm around the complex and can be shown to be displaced sinistrally by some 6km across the Strathconan Fault (Hutton *et al.* 1993).

Regionally, the Microdiorite Suite dykes are late- to post-regional ductile deformation, whilst the Minette Suite entirely post-date the regional deformation (Smith 1979). Direct evidence of age relations is found north of the complex on the Shore of Loch Duich (Map 3 NC 892 228) where a 1m wide microdiorite is cross-cut by a thick minette dyke.

5.3.5 *Late veins*

The quartz monzonite is cut by quartz-fluorite-calcite veins, particularly in the vicinity of the Strathconan Fault system (Alderton 1988). The ore mineralogy of these veins is dominated by pyrite, chalcopyrite, galena and sphalerite, with minor hessite and electrum. Gold occurs in the electrum and silver in the galena (Alderton 1986, 1988). The high temperatures indicated by fluid inclusions and the enrichment of the hydrothermal fluids in Sr and Ba suggests a close link between this mineralisation and magmatism.

5.3.6 Compositional features

The major oxide composition of the complex indicates that it is enriched in alkalis, especially Na_2O , when compared with the entire "Newer Granite" province (c.f. Nicholls 1951a,b). The complex as a whole describes a smooth trend on a $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{FeO} + \text{Fe}_2\text{O}_3 - \text{MgO}$ plot, typical of calcalkaline igneous series (Stephens & Halliday 1984). Trace element studies (Halliday *et al.* 1984) indicate notably high values of Sr (typically 2000-5000ppm in the diorite, reflecting primary celestine) Ba (up to 6000ppm in the syenitic facies) and Ce (up to 500ppm in the diorite, reflecting high titanite abundances). The early gabbros and pyroxene-mica diorites have the following general isotopic composition (Halliday *et al.* 1984):

Pyroxene mica diorite and gabbro = $(^{87}\text{Sr}/^{86}\text{Sr})_{425} \sim 0.7058$, $\epsilon_{\text{Nd}} \sim -4$, $\delta\text{O}_{18} \sim -7\%$

Amphibole diorites and monzonites = $(^{87}\text{Sr}/^{86}\text{Sr})_{425} \sim 0.7052$, $\epsilon_{\text{Nd}} \sim -13$, $\delta\text{O}_{18} \sim 7 - 9\%$

Some appinites = $(^{87}\text{Sr}/^{86}\text{Sr})_{425} \sim 0.7058$, $\epsilon_{\text{Nd}} \sim -7$, $\delta\text{O}_{18} \sim 6 - 8\%$

According to Hutton *et al.* (1993), these data are consistent with the likely parental magmas for the complex (gabbro and pyroxene mica diorite) being relatively primitive melts, since they have 'mantle-like' signatures in trace elements and isotopes.

5.4 The deformation history of the plutonic complex

The deformation history of the Ratagain Complex can be generally subdivided into early and late episodes. However, in detail the history of deformation varies between the dioritic and monzonitic units, with the quartz monzonite suffering some high temperature solid state deformation which appears to be absent in the diorite. The relevance of these slight variations will be discussed later in this section.

5.4.1 Early deformation

A penetrative, steeply inclined magmatic state deformation fabric of varying intensity is found throughout the pluton (Fig. 5.3; Hutton & McErlan 1991). It is a flattening fabric, with a weak

maximum elongation direction, that is S or LS in the nomenclature of Flinn (1965). Where preserved, the linear component is sub-horizontal or gently plunging. In outcrop and hand specimen, the fabric varies in appearance from a crude alignment of feldspar and hornblende phenocrysts and mafic clusters to a well-developed and obvious planar anisotropy of these phases.

(a) *Early deformation in the diorite*

In outcrop and hand specimen, the early deformation fabric is as described above, and is a variably developed magmatic state deformation fabric. This is confirmed in thin section, where plagioclase and hornblende phenocrysts are typically euhedral, aligned and show no evidence of any internal lattice distortion (Plate 5.2). Biotite occurs in elongate clusters and shows no sign of having been plastically deformed. Groundmass minerals are equant, subhedral and show only minimal amounts of internal lattice distortion, such as weakly developed undulose extinction in quartz (Plate 5.3). Most of the deformation is interpreted to have occurred before the magma had reached its rheologically critical melt percentage (RCMP), and therefore before it had completely crystallised.

(b) *Early deformation in the quartz monzonite*

The quartz monzonite exhibits a broadly similar fabric to the diorite, with alkali feldspar, plagioclase and hornblende phenocrysts are aligned in a relatively undeformed groundmass of quartz and biotite clusters. In addition, however, the monzonites show some evidence of a high temperature solid state overprint to the magmatic state fabric. This involves limited development of subgrains and new grains around the margins of feldspar (plate 5.4) and, on rare occasions, hornblende phenocrysts. In addition, deformation-induced myrmekite (Simpson 1985) has developed at points of higher normal stresses around alkali feldspar phenocrysts (Plate 5.5), plagioclase twin planes may be offset by microscopic healed shears and, in extreme cases, biotite may be kinked. None of these features are pervasively developed and it is the view of the present author that they could be produced by a slight elevation in the strain rate associated with the magma reaching its rheologically critical melt percentage, at which point, phenocrysts would start to impinge on one another. However, why they are focused into

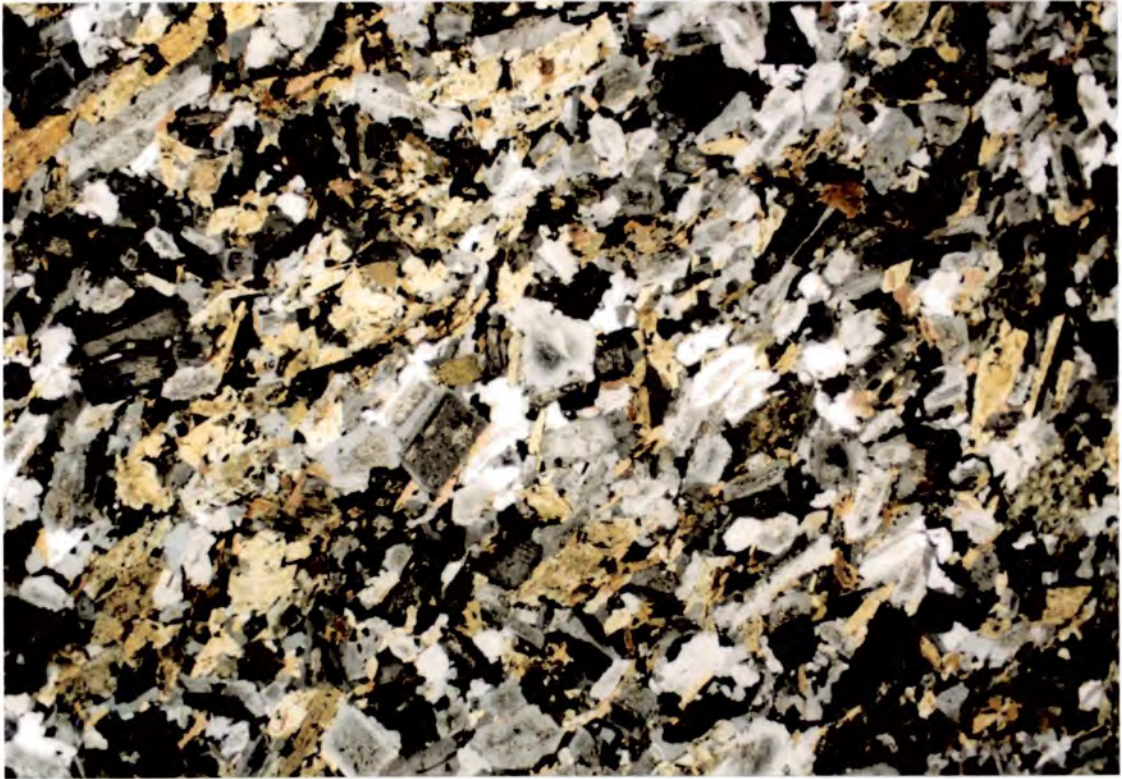


Plate 5.2 Photomicrograph of a magmatic state deformation fabric in a specimen of diorite from the Ratagain Complex. The magmatic state fabric is defined by a well-developed preferred orientation of phenocryst phases, set in a groundmass of equant quartz. (Field of view 18mm).



Plate 5.3 Example of very minor amounts of plastic strain within groundmass quartz, exhibited as undulose extinction, in a specimen of diorite that carries a predominantly magmatic state fabric. (Field of view 1.9mm).

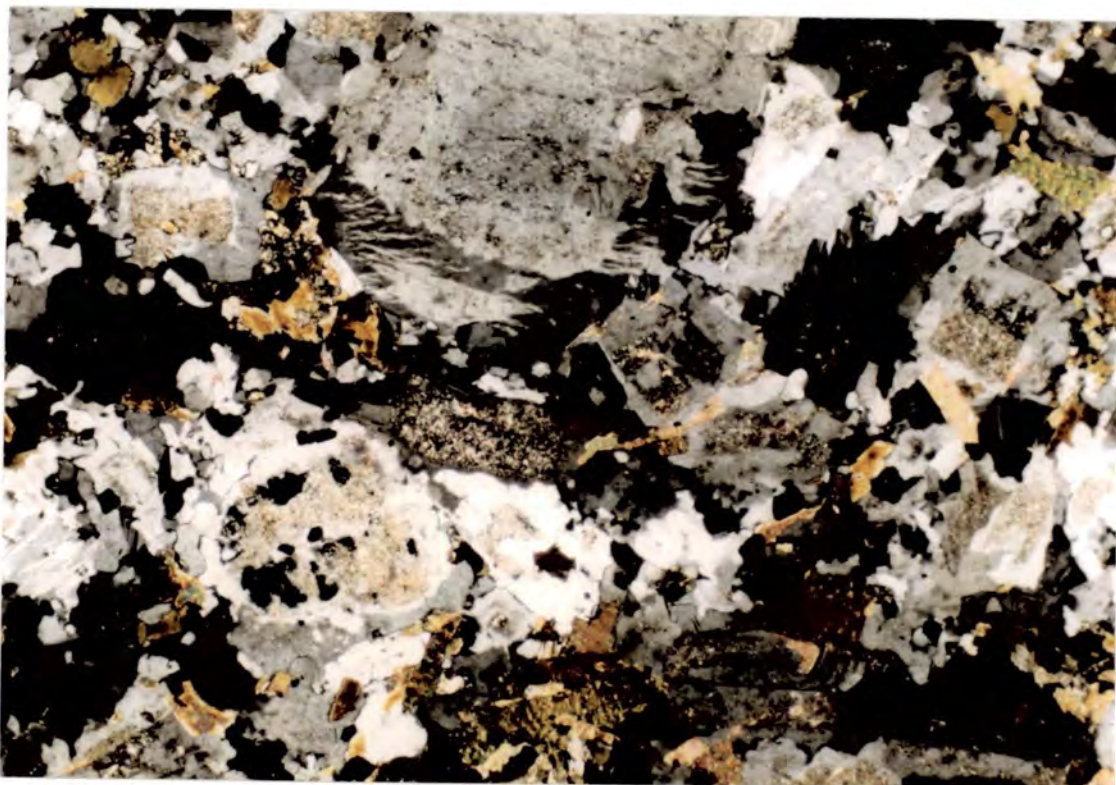


Plate 5.4 Photomicrograph showing high temperature solid state deformation fabric overprinting a well-developed magmatic state deformation fabric in a sample of monzonite from the Ratagain Complex. The solid state deformation is relatively weak and results in the development of subgrains around the margins of feldspar phenocrysts and limited recrystallisation of groundmass quartz. (Field of view 4.4mm).

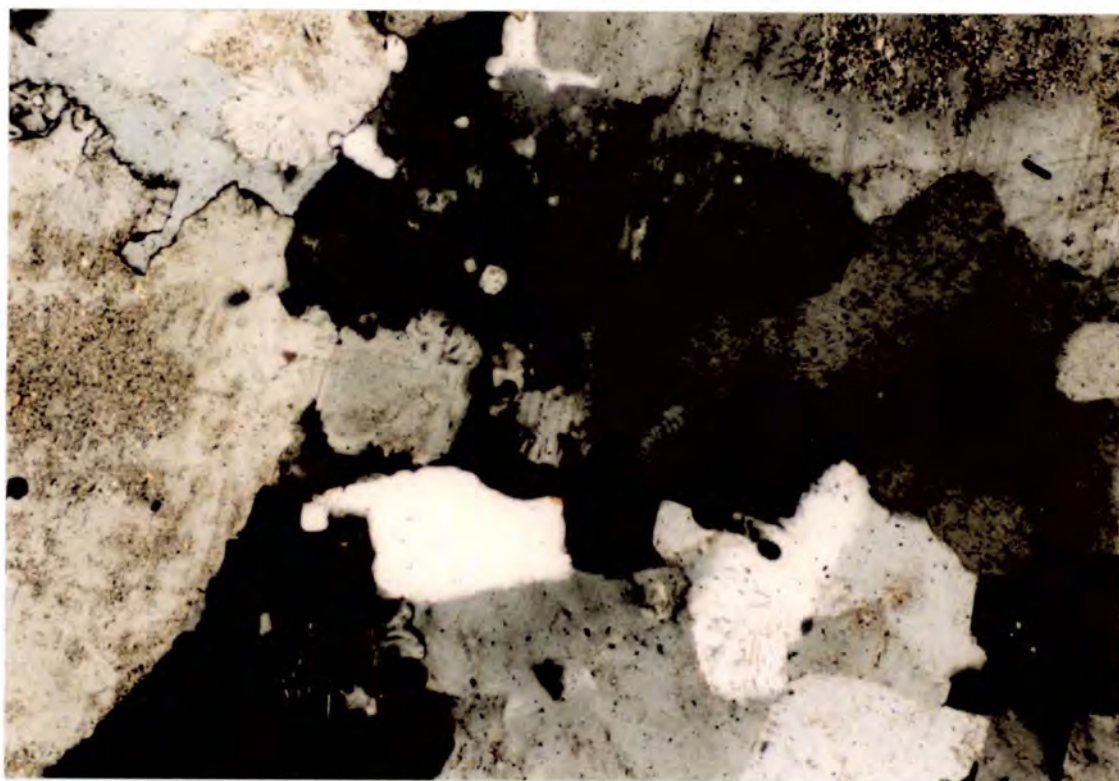


Plate 5.5 Photomicrograph illustrating the presence of strain-induced myrmekite as new grains at the margins of alkali feldspar phenocrysts. (Field of view 1.9mm).

the monzonite unit and not generally in the slightly earlier diorites is unclear. One possibility is that the locus of deformation moved into the younger and more fluid monzonite when that body was emplaced and that is where the last waning stages of the early deformation occurred; this is a form of strain partitioning, where deformation localises itself in the 'softer', more easily deformable zone. Hutton & McErlan (1991) suggested that another possible explanation for the difference in behaviour between the diorite and the quartz monzonite is that solid state deformation could have started earlier in the monzonite because higher vapour pressures, for which there is evidence (Hutton *et al.* 1993) in this unit, resulted in lowering of the solidus and caused earlier onset of crystallisation compared to the adjacent diorites. However, this author now believes this hypothesis to be incorrect, since lowering the solidus would not result in the earlier onset of crystallisation, but would tend to extend the magmatic interval, in particular, the interval between the RCMP and complete crystallisation, and it is the latter that would tend to promote the development of high temperature solid state fabrics. Whatever the reason, the amount of strain associated with the high temperature modification in the monzonites is minor when compared with that associated with the magmatic state deformation fabric in these rocks. It is therefore the conclusion of this author and the other investigators in this project, that the bulk of the early deformation in the Ratagain Complex occurred during the crystallisation interval of the component magmas, and had generally ended before crystallisation was complete.

5.4.2 *The origin of the early deformation fabric*

The origin of the early fabric can be considered by examining macroscopic and mesoscopic features of its geometry and kinematics. In the diorites and appinites in the south and west of the pluton, within and west of the Glenmore river, the steep fabric swings in trend from NW-SE to NE-SW moving southeastwards (Fig. 5.3). In the river section, where strikes are NE-SW, deformed microdiorite xenoliths consistently indicate a sub-horizontal direction of maximum extension. In the northern part of this section, X/Z strains (measured using the methods described in Hutton 1982 & 1988a) remain generally low (values of approximately 3.0). However, on moving southwards towards the granite contact nearest to the Strathconan Fault, they rise steadily over an across-strike distance of 1km to values of approximately 6.0.

The increase follows an LS-S deformation path, with a K value of approximately 0.4, that is, within the flattening field (Fig.5.7). In the zone of higher strain dioritic/granodioritic banding is well developed and in one locality within the zone of highest strain, the banding and the magmatic state deformation fabrics are deflected and offset in a sinistral sense across a healed shear 0.5cm wide and at least 2m long. In thin section the healed shear contains fine-grained, euhedral quartz, with approximately 120° triple junctions, and small microclines which have suffered considerable ductile lattice distortion (Plate 5.6). In the immediate wall rocks to the shear zone, strong magmatic state fabrics show much evidence of high temperature solid state modification, with distortion and kinking of feldspars, often along twin planes, and flattening of quartz into lensoid shaped aggregates. These features are consistent with what Ingram (1992) has termed 'lock-up shears', meaning that they occurred around the RCMP, when constituent crystals in the magma start to impinge on one another.

FLINN DIAGRAM ILLUSTRATING FLATTENING STRAINS
IN A SECTION ALONG THE GLENMORE RIVER

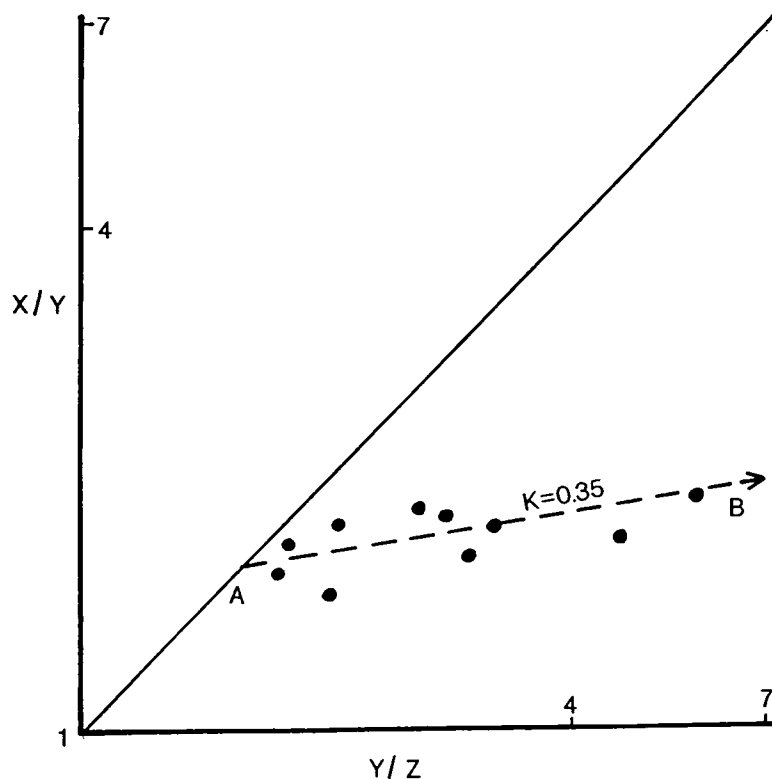


Figure 5.7 Flinn diagram to illustrate flattening strains in diorite exposed in the Glenmore River.

It is thought that the general swing in the strike of the fabric, the sub-horizontal maximum extension direction, the zone of steadily increasing strain and the small, but important, healed shear are consistent with the presence of a major NE-SW trending sinistral shear zone at the southeast margin of the pluton. The microscopic data suggest that the shear zone was in motion *before* the plutonic units had completely crystallised. As the temperature dropped and crystallisation proceeded towards completion, the deformation became concentrated into discrete zones of high strain, generating highly localised, high temperature solid state deformation fabrics. These zones subsequently annealed at temperatures in excess of 450°C, since ductile deformation of feldspar generally does not occur below this temperature (Simpson 1985). Such an evolution of fabrics within a cooling and deforming granitoid exactly follows the sequence predicted by Gapais (1989).

The general swing in the fabrics from NW-SE to NE-SW on moving southwards is also seen in the monzonite unit in a well exposed traverse along a road cutting and in the shores of Loch Duich (Figs. 5.3 & 5.8). The sections in the Glenmore River and by Loch Duich come closest to the trace of the brittle Strathconan Fault. Between these two sections the country rock contact protrudes northwestwards into the pluton. This contact is part of the gently dipping roof, and it is considered likely that the NW trending fabrics immediately adjacent to this contact would also swing counterclockwise beneath the roof segment towards parallelism with the trace of the fault. The observed data suggest, therefore, that the locus of the sinistral shear zone, operating during and immediately after the emplacement of the Ratagain pluton, lay close to or along the current trace of the Strathconan Fault (Fig.5.8; see also Hutton & Mc Erlean 1991).

The general fabric pattern appears to be modified by another, narrower zone, which crosses the centre of the pluton from WNW to ESE. This deflects the overall fabric pattern with a sinistral sense (Figs. 5.3 & 5.8) and relatively more intense strains are observed along its axis. There is also smaller scale evidence of sinistral deflections in this area, where fabrics in the diorite are displaced anticlockwise against the margins of metre scale xenoliths of country rock. This zone is interpreted as a relatively high angle synthetic splay to the main shear zone, which lies to the SE (Fig. 5.5). Another parallel splay may exist in poorly exposed ground to the NE (Figs. 5.3 & 5.8; see also the emplacement model outlined in Fig. 5.13; Hutton & Mc Erlean 1991).

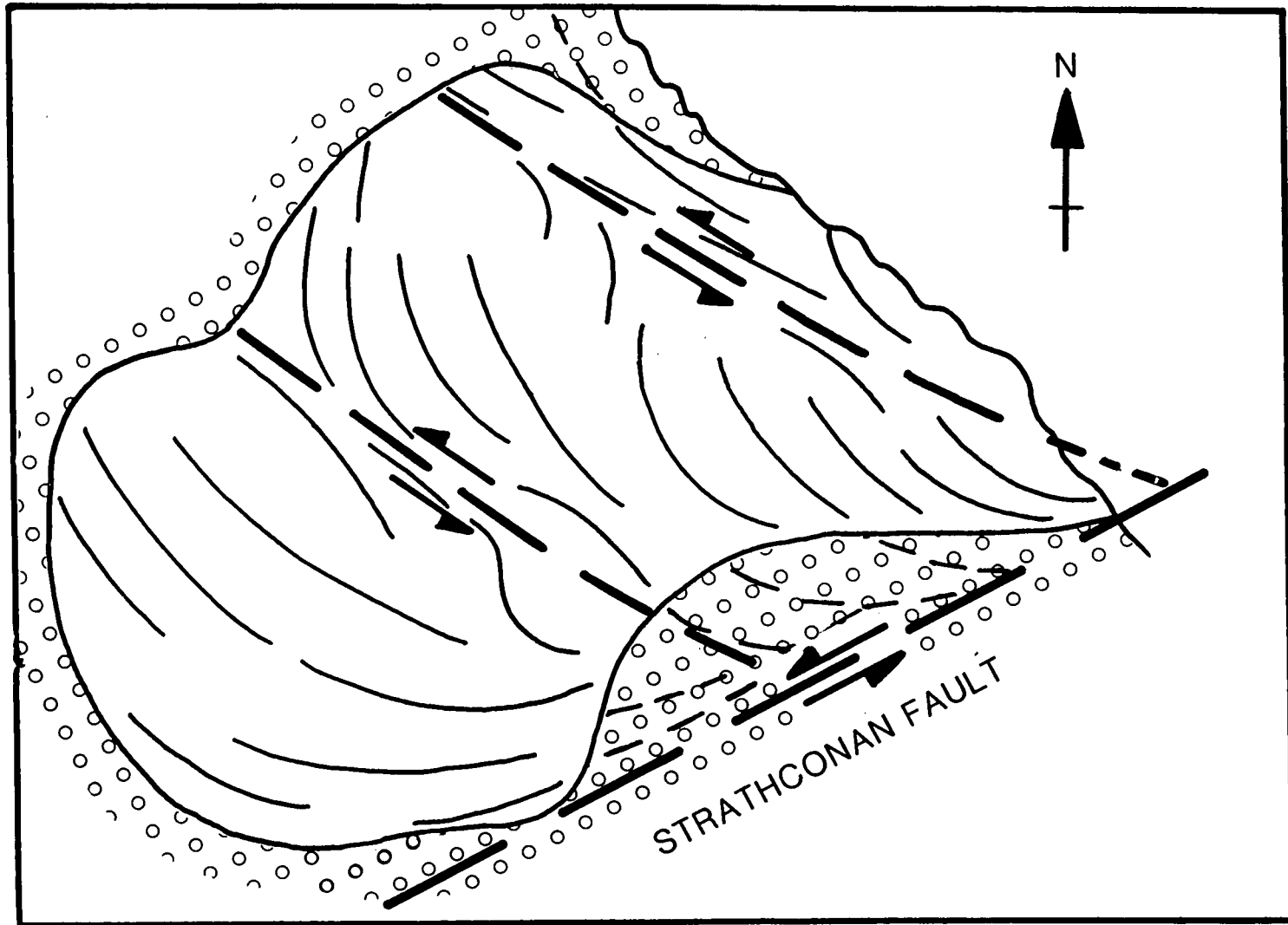


Figure 5.8 Magmatic state foliation trends within the Ratagain Complex and the position of the putative shear zones at the time of emplacement and deformation.

In section 5.5 the data presented here have been interpreted as being consistent with magma emplacement into a region of complex transtension deformation regime adjacent to the Strathconan Fault. Since this early deformation had ended before the complete crystallisation of the magma, the age of this deformation event is closely approximated by the crystallisation of the pluton at 425 ± 3 Ma (Wenlock) (Rogers & Dunning 1989).

5.4.3 *Late deformation*

The Ratagain Complex is also deformed by a network of late-stage, low temperature shear zones and brittle faults on a variety of scales from metres to kilometres in length. Within these, planar, relatively low temperature mylonites, with sub-horizontal stretching lineations, and cataclasites and breccias, associated with sub-horizontal shear fibres and striations, are extensively developed. The mylonitic rocks in both the diorite and monzonite contain porphyroclasts of feldspar and mica, set in a fine-grained ($5\mu\text{m}$), dynamically recovered and recrystallised matrix of equant quartz grains (Plate 5.7). Mylonites with more micaceous matrices often show strong foliations, with shear bands being common in the most phyllonitic horizons.

Chlorite is the only stable mafic phase and this, together with the dynamically recovered nature of many of the quartz groundmass textures, suggests that the temperatures during the late deformation event approximated to that of the low greenschist facies (Simpson 1985; Gapais 1989). Breccias and cataclasites often locally cross-cut mylonites, suggesting that they are later and possibly associated with lower temperature conditions during deformation, since the mechanisms of deformation appear to have crossed the brittle-ductile transition (Simpson 1985; Gapais 1989). In mapping the pluton, Hutton directly observed over 200 small-scale faults and shear zones with determinable shear sense were measured in 5 sub-areas. Mylonite zones show broadly similar orientation patterns to breccia and cataclasite zones, so Hutton analysed both together. Results of these analyses are summarised in Hutton and McErlean (1991), who reported that in each sub-area, second order synthetic and antithetic orientations indicated a maximum principal stress orientation counterclockwise and oblique to the first order NE-SW orientations. This, together with the directly observed shear sense on the first order faults

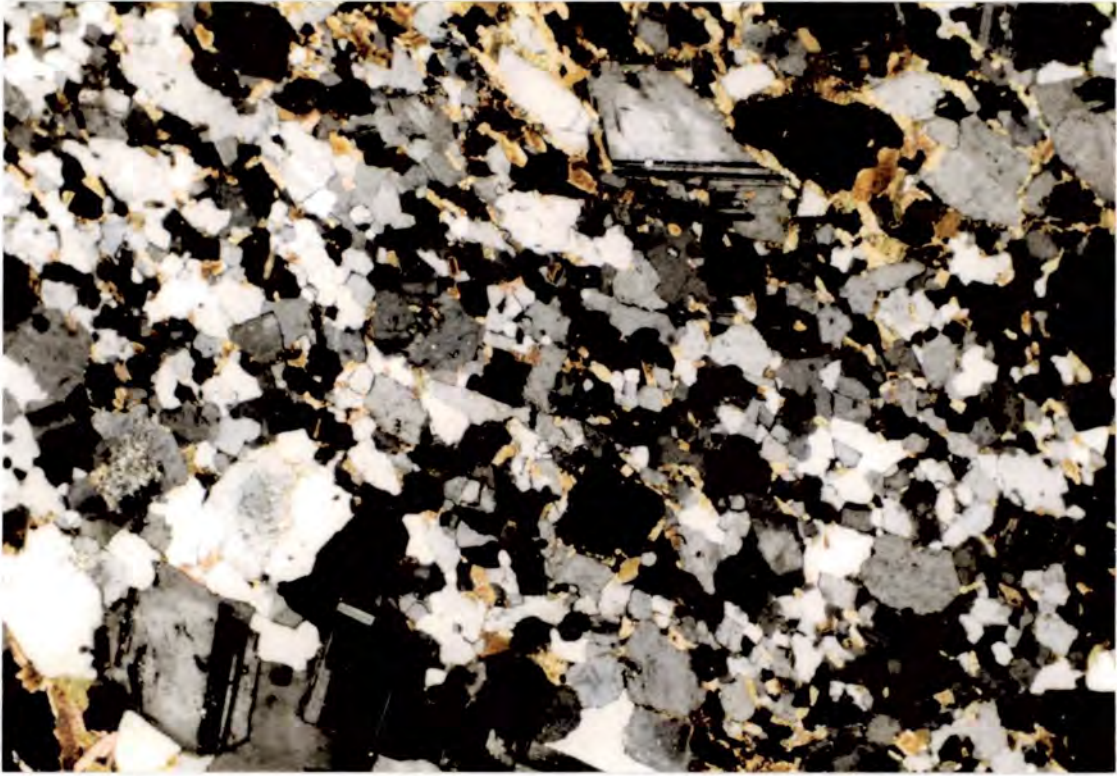


Plate 5.6 Photomicrograph of an annealed texture within a healed shear zone in diorite specimen collected from the Glenmore River, close to the southern contact of the complex. Note the polygonal arrangement of quartz grain boundaries; this is characteristic of annealing. (Field of view 4.4mm).



Plate 5.7 Photomicrograph of mylonitised diorite from a discrete zone containing porphyroclasts of feldspar and mica set in a fine grained, dynamically recrystallised matrix of relatively equant quartz grains. (Field of view 18mm).

suggest an overall sinistral pattern (Hutton & McErlean). A lineament analysis of the major tectonic features in the pluton and its immediate vicinity is consistent with this pattern (see Hutton & McErlean 1991).

Dykes and steeply inclined sheets of porphyrite, felsite and minette occur commonly in this area. They cut the plutonic complex in a WNW-ESE direction in two main swarms (Fig.5.9). These dykes entirely *post-date* the early deformation, cutting across the magmatic state fabrics in the main complex, but they both cut and are cut by the cataclastic and breccia zones. This suggests that they are penecontemporaneous with the later deformation. Both swarms show internal, sinistral, en echelon patterns (Hutton & McErlean 1991; Hutton *et al.* 1993). In addition, the southern swarm appears to be coincident with the central part of the early WNW-ESE trending sinistral splay associated with the early deformation. This may also be the case with the less well exposed northern swarm and splay. In the light of these data, it is therefore proposed that the splays, possibly as basement faults of the Loch Maree - Kinloughhoun fault set, were partially reactivated during the late deformation to control dyke location and orientation, rather than the latter being directly controlled by movements along the Strathconan Fault.

Rock (1988) reported that the regional minette dykes of the Highlands have been dated at 410-395 Ma by Rb-Sr techniques. At Ratagain the swarm is offset sinistrally across the Strathconan Fault and yet locally, as mentioned before, cuts breccias associated with this fault. These data indicate that the late sinistral deformation of the Ratagain pluton was associated with movement on the Strathconan Fault during the intrusion interval of the regional minette dyke swarm, at 410-395 Ma (Late Silurian - Early Devonian).

In addition to the offset of minor intrusions, there is the occurrence of the Glen Licht body southeast of the Strathconan Fault. If this was originally a part of the complex, as suggested by Dhonau (1964), then its current position indicates that there has been a post-granite emplacement and cooling displacement of approximately 6km along the Strathconan Fault.

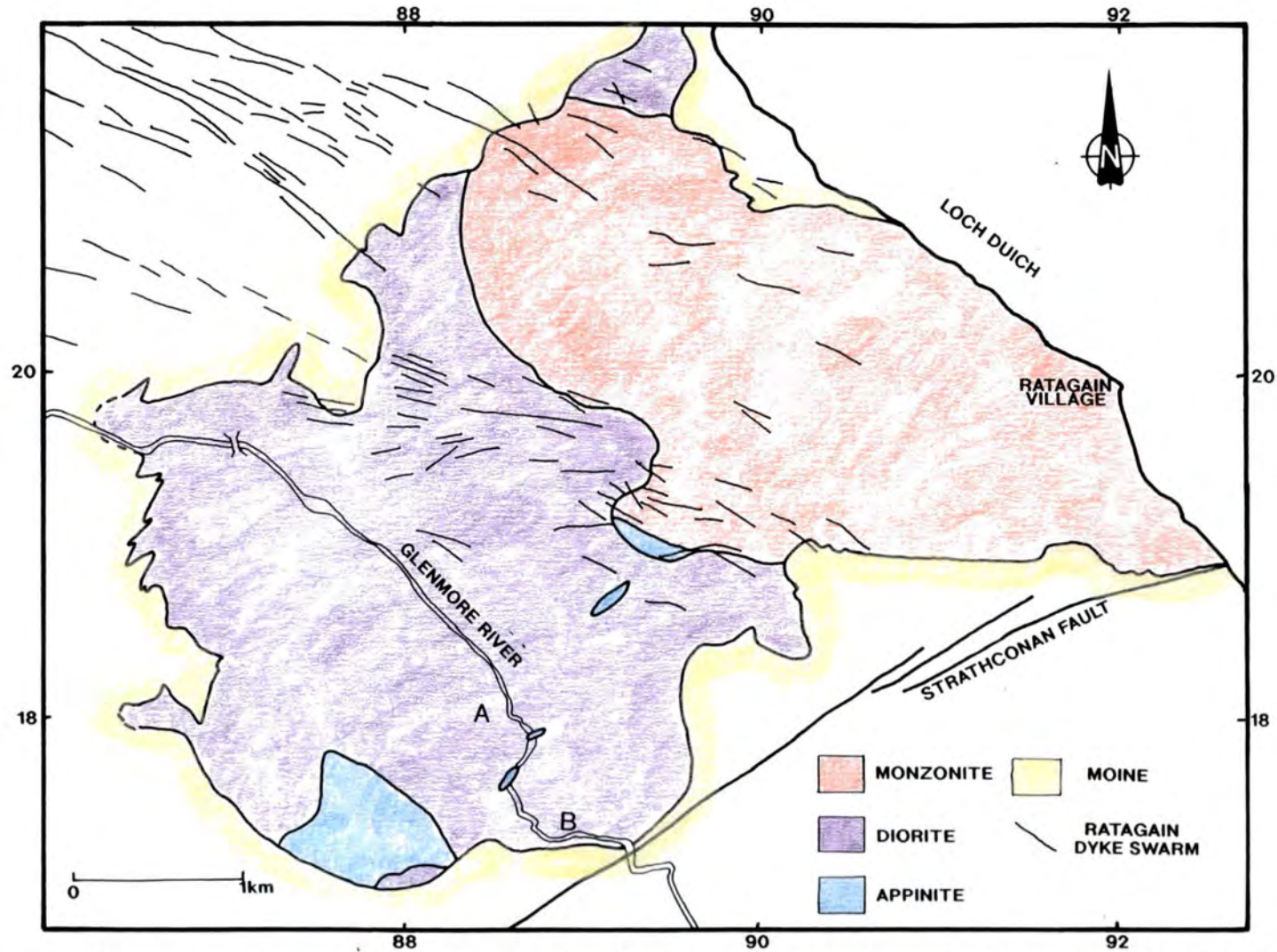


Figure 5.9 Map illustrating the major intrusive units of the Ratagain Complex and the location and orientation of Lamprophyre and felsite dykes.

5.5 Emplacement model for the Ratagain Igneous Complex

The mode of emplacement of the Ratagain complex has either been ignored by or has defied previous investigators. Deducing a model of emplacement is made difficult by the fact that the pluton contacts are often very poorly exposed. Other enigmatic features include the fact that the pluton appears to have a most peculiar shape, with the field relationships suggesting that the present exposure level is very close to the apparently flat roof of the pluton, whilst interpretation of geophysical surveys (gravity and aeromagnetic) suggest that the pluton has steep sides at depth. The cartoon sketches shown in Fig. 5.10, attempts for the first time to interpret the three dimensional form of the pluton based on all available data. In addition, as the pluton has, on first glance, a pseudo-concentric appearance in the field, this might suggest a diapiric, or ballooned pluton geometry. However, this not believed to be true, since strain and other data from within the pluton are not consistent with diapiric or ballooning intrusions; the strains in the pluton *do not* generally increase concentrically towards the pluton margins; the country rocks *are not* deflected parallel with the outer contact of the pluton, *nor* do they show any increase in strain intensity towards the plutonic complex. The fact that some of the magmatic state deformation fabrics are parallel to the pluton walls could perhaps be attributes to 'buoyancy head' forces developing in response to the magma being confined within relatively rigid walls (see Ch. 4).

The pluton's flat roof could perhaps be explained by invoking synchronous movements on the Strathconan fault and the Moine Thrust. However, this is not believed to be feasible in terms of the regional deformation chronology, which suggests that compressional movements on the Moine Thrust (c. 435Ma.; Johnson *et al.* 1985) had ceased *before* the initiation of transcurrent movements on the Great Glen Fault system, to which the Strathconan Fault belongs (Watson 1984). At the presently exposed structural level, the magmatic fabrics are predominantly steeply dipping and appear to be associated with strike-slip displacements, making it unnecessary to specifically invoke involvement with the Moine Thrust, or even late low angle extensional features which may have developed at a later stage within the Moine Thrust Zone (Coward 1983). Rather, the data presented in section 5.4 are believed to be consistent with magma emplacement in association with transtension which resulted in the creation of a cavity into

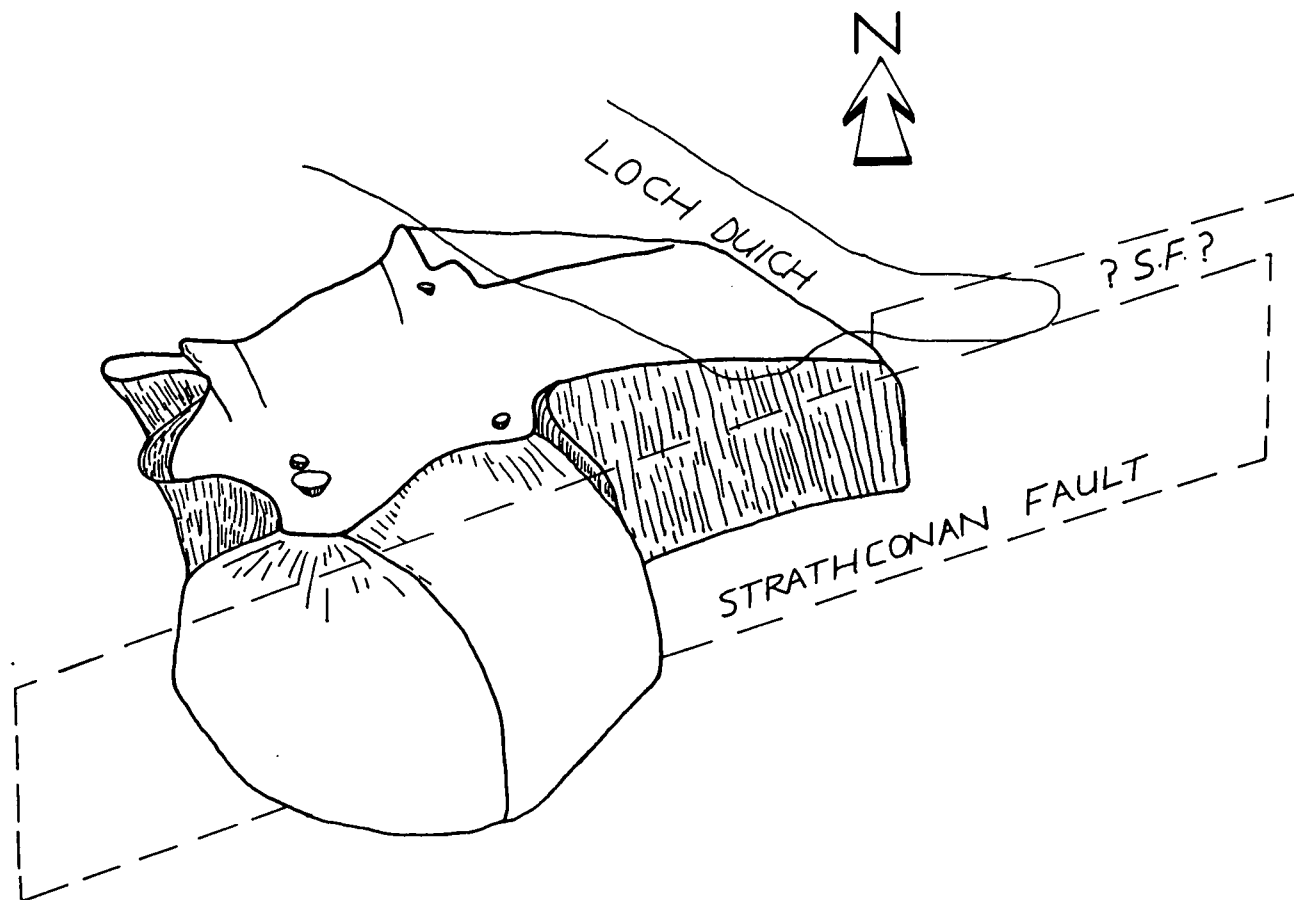


Figure 5.10 Cartoon diagram illustrating the general shape of the Ratagain Complex as derived from field mapping and interpretation of geophysical data (gravity & aeromagnetic). Includes present day position of the Strathconan Fault.

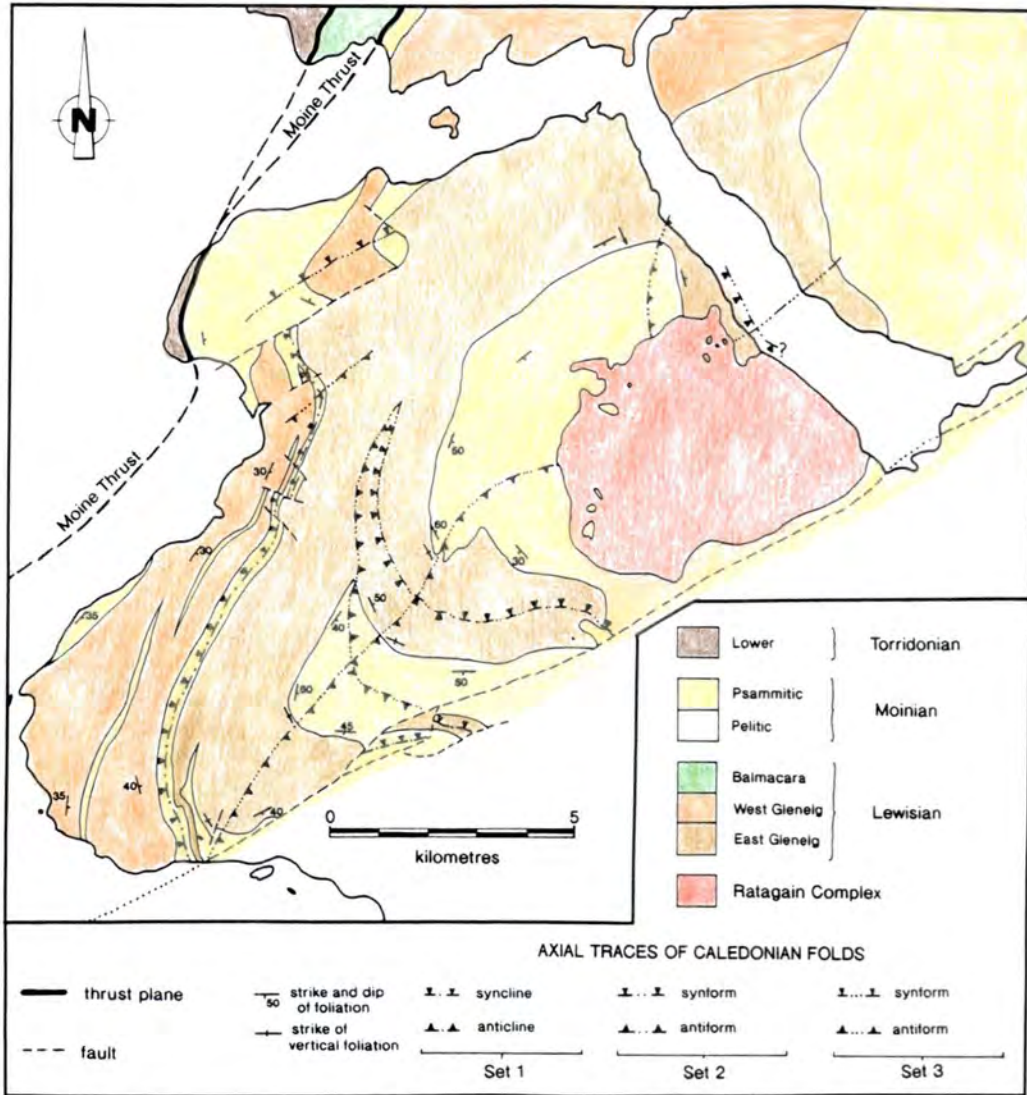


Figure 5.11 Map illustrating the major stratigraphic units and the main folds and faults in the Glenelg area.

which magma could be emplaced. The exact manner in which this would have occurred might depend on the exact geometry of the Strathconan Fault/shear zone immediately prior to granite intrusion. The fault as it is seen today does not appear to show any significant bends in the area of the Ratagain Complex which would have resulted in the development of, for example, pull-aparts in relation to releasing bends or offsets, or similar geometries which have been described in the literature to explain granite emplacement in response to tectonism (Castro 1986, Sibson 1987, Hutton 1988). This does not necessarily mean that these geometries did not exist at the time of pluton emplacement, but, without definite present-day evidence for such features, it is not possible to invoke their involvement in the emplacement of the Ratagain Complex.

The next step in deducing an emplacement model is to re-examine what we know about both the pluton geometry, the form of the country rocks and their part in the intrusion process and the kinematics of intrusion. On close inspection of the published maps of the area (Sutton & Watson 1958; Ramsay 1958; Sanders 1979; BGS 1" series, Kintail sheet & Kyle of Lochalsh sheet) and the maps drawn up from this study (Map 3 & Fig. 5.11) one can see that there is a general parallelism between the country rock lithologies of the Glenelg area and the contacts of the Ratagain Complex. Several complex deformation events are responsible for the present day disposition of the country rocks in the Glenelg area (as described in section 5.1). However, possibly the most influential factor is the geometry of the various large-scale fold phases seen in the area, the latest of which appear to be F_3 structures (Sutton & Watson 1958; Ramsay 1958 [he called these F_2 structures]). The most significant feature is that the trace of a major F_3 structure, the Beinn a' Chaonnich antiform, runs through the middle of the igneous complex (Fig. 5.11). In the following section, I present a simple intrusion mechanism for the pluton that involves interaction between the folded structure of the country rocks and sinistral movements on the Strathconan Fault in the presence of magma at depth (see fig 5.12 and following text).

5.5.1 *The emplacement model*

Figure 5.12 is a cartoon sketch which summarises the emplacement model for the Ratagain Complex in stages.

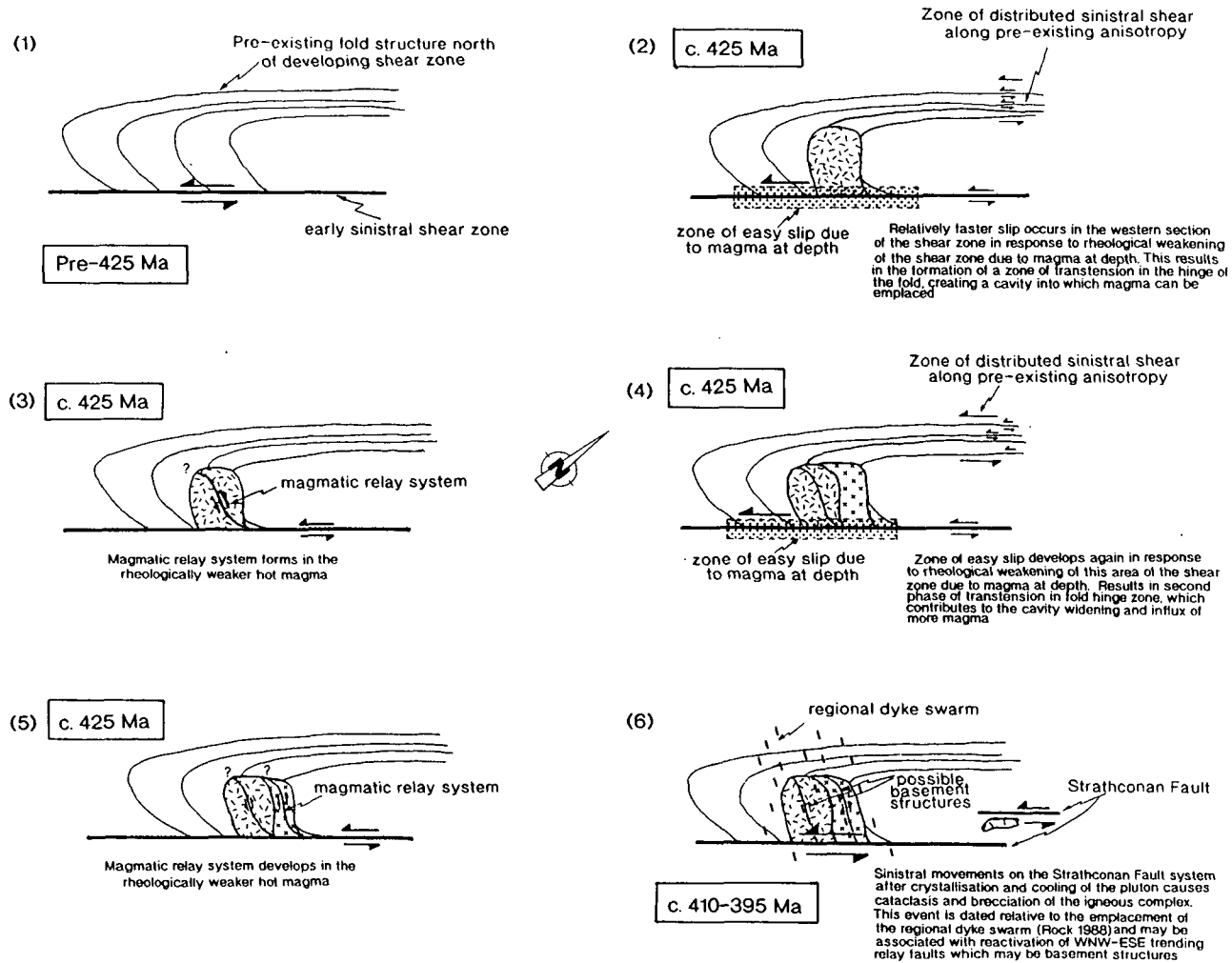


Figure 5.12 Sketch diagram illustrating the proposed stages of emplacement of the Ratagain Pluton.

Stage 1: Involves the development of the an early shear zone close to, if not at, the present day trace of the Strathconan Fault. Early movements along this shear zone were probably associated with ductile deformation, and initiated somewhere between 430 and 425 Ma according to published chronologies (Watson 1984). The shear zone lies to the south of a major F_3 fold, the Beinn a' Chaonnich antiform, which belongs to a set of structures which may be associated with, or at least influenced by, the earliest ductile movements on the Strathconan Fault (Ramsay 1958; Sutton & Watson 1958).

Stage 2: Sinistral movement along the Strathconan shear zone continues, with a zone of relatively easy slip developing towards the SW end of the shear zone, in response to rheological weakening of this part of the zone due to the presence of magma at depth (see Ch. 7). As a result, the NE end of the shear moves at a relatively slower rate than the SW end, producing a zone of *transtension* around the hinge zone and the SE limb of the pre-existing F_3 fold structure, resulting in the production of a cavity. Transtension focuses in these areas because it is here that the dominant anisotropy is at high angles to the shear zone. A first pulse of magma, the diorite unit, then invaded the cavity. This must have happened in a fairly passive manner (c.f. Ch. 1) since there are no obvious emplacement related deformation structures in the surrounding country rocks. There may have been a small amount of stoping involved with magma intrusion, since xenoliths of country rock, interpreted to be roof pendants, are found within the intrusive units. Such a model would require sinistral shearing, for which there is some evidence (Sutton & Watson 1958) to the N and NW of the pluton. The present model has accommodated this necessary sinistral shearing along the pre-existing anisotropy, where it is sub-parallel with the main shear zone in the NE limb of the Beinn a' Chaonnich antiform.

Stage 3: Sinistral shearing along the main shear zone continued during diorite emplacement, resulting in the development of magmatic state deformation fabrics within the dioritic unit and the production of relatively higher strains (as measured from deformed cognate xenoliths; see sub-section 5.4.2) in the south of the intrusion, nearest the trace of the present day Strathconan Fault. This simple fabric pattern is modified by the presence of a WNW-ESE trending high strain zone running through the pluton. This deflects the magmatic state deformation fabric with anticlockwise sense of motion, consistent with sinistral shearing, and is also associated with slightly higher strains. An explanation for this high

strain zone is that the presence of hot magma within the hinge and SE limb of the early fold structure, facilitating the development of a synthetic splay through this rheologically weaker material. It is uncertain where the displacement along the splay goes when it reaches the other side of the magma body, but one explanation for this is that the deformation is simply dissipated within the zone of distributed sinistral shearing along, or parallel to the pre-existing anisotropy.

Stage 4: This basically involves a repetition of stage 2 (see Fig. 5.12) in response to the arrival of another pulse of magma at depth. The magma 'softens' the SW end of the shear zone, resulting in the development of another zone of transtension which permits opening of a cavity, allowing the quartz monzonite to be emplaced. As in stage 2, sinistral shearing to the N and NE of the pluton is accommodated along the pre-existing country rock anisotropy in the NE limb of the Beinn a' Chaonnich antiform.

Stage 5: A second magmatic relay systems, with a sense of movement synthetic to the main shear zone, develops within the hot quartz monzonite magma (Fig. 5.12), producing a deflection of the magmatic state deformation fabric in a WNW-ESE trending zone of relatively high strain. Deformation at this stage of pluton development tends to focus in the slightly later and hotter quartz monzonite magma, since rheologically this is the 'softest' material, hence it is easiest to deform. This would explain why solid state deformation fabrics are absent from the dioritic unit, even though deformation continued during the emplacement of the slightly later quartz monzonite unit. The early deformation event must have ended soon after the intrusion and partial crystallisation of the quartz monzonite magma, so that only magmatic state and high temperature solid state deformation fabrics are seen in the intrusive units, with pervasive solid state deformation fabrics being almost entirely absent from the pluton.

Stage 6: Involves post-crystallisation and low temperature deformation of the of the pluton in response to renewed movements along the Strathconan Fault. This deformation event resulted in the development of cataclastic and breccia zones within the pluton, which Hutton *et al.* (1993) indicate as having geometries consistent with sinistral shearing. In addition to producing cataclastic zones within the pluton, this deformation event is responsible for the present outcrop pattern around Loch Duich, where small outcrops of heavily brecciated diorite and quartz monzonite can be observed, they may also account

for the separation of the Glen Licht body from the main igneous intrusion (Dhonau 1964). This deformation event spans the intrusion interval of the regional Minette dyke swarm (410-395 Ma Rock 1988) since the dykes both cut and are cut by the Strathconan Fault, with some of the suite exhibiting definite sinistral deflections across the fault (Watson 1984; Hutton & McErlean 1991). The intrusion of the regional dyke swarm within the igneous complex appears to follow the WNW-ESE trending relay shear zones which, in turn, may follow basement structures at depth. This emphasises the importance of pre-existing anisotropy in the country rocks; if they were isotropic, sinistral shear along the Strathconan Fault should produce approximately N-S trending dykes (c.f. Wilcox *et al.* 1973). Such simplistic model are clearly inappropriate in basement terrains of this kind. It is important to stress that, in contrast to the argument presented by Watson (1984), the present model requires that the latest deformation of the pluton related to sinistral movements on the Strathconan Fault is a completely separate phase of movement to the early events related to granite emplacement. Were they part of a single movement event, one would expect to see the development of pervasive solid state deformation fabrics within the pluton, bridging the gap between magmatic state and cataclastic deformation fabrics.

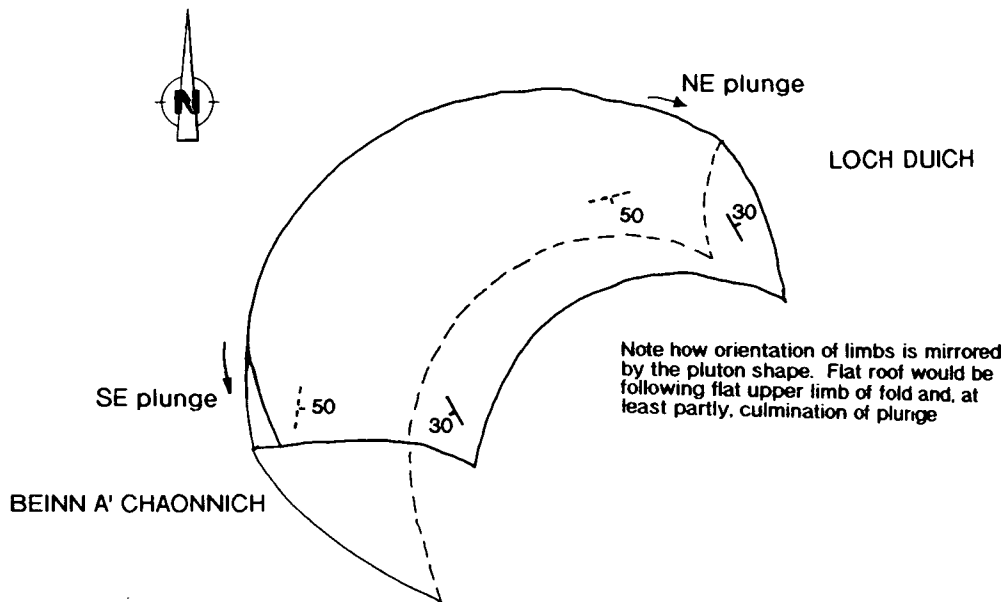
One assumption of the model outlined above is that the magma played a fairly passive role in terms of opening the cavity into which it was emplaced. This assumption has been made since there is little or no evidence of forceful emplacement in the country rocks surrounding the pluton, although it is not possible to assess the degree of forceful activity beneath the present level of exposure, which is believed to be close to the roof of the pluton. If one assumes that the magma played little or no part in the opening of the cavity, then it seems feasible to envisage the magma being drawn into the open cavity in a manner analogous to the 'suction pump' behaviour of fluids proposed by Sibson (1981, 1987) to explain the fluxes of hydrothermal fluids in zones of active faulting. This model suggests that fluids will be drawn into zones of dilation where the mean stress is significantly reduced subsequent to rapid displacement episodes along the main shear zone.

5.5.2 Discussion of the emplacement model

The suggested model is a suitable explanation for the deformation fabrics and geometries produced in association with granite emplacement and it seems to fit the geometry of the pluton, as schematically portrayed in Figure 5.12. Further support for the model lies in the interpretations of the deformation chronology for the Glenelg area, and more generally for Inverness- and Ross-shire. To briefly recap on the deformation history and chronology, the Moine and Lewisian were interbanded at an early stage by isoclinal folding (Clough 1910; Ramsay 1958; Sutton & Watson 1958). These interleaved slices of Moine and Lewisian were then tightly refolded by structures which plunged mostly to the E and SE. These folds were called F_1 structures by Ramsay (1958) because he failed to recognise any earlier fold structures around the Glenelg area, but were correctly identified as F_2 structures by Sutton & Watson (1958). Ramsay (1958) showed that these F_2 folds, such as the Beinn a' Chapuill antiform and the Gleann Beag synform, were refolded by yet another fold phase, F_3 in the current chronology but F_2 according to Ramsay. The largest of the F_3 folds in the Glenelg area is the Beinn a' Chaonnich antiform, which runs through the middle of the Ratagain Complex. It can be correlated in terms of age and geometry with the Beinn Sgriol synform (Ramsay 1960), which lies to the south of the Strathconan Fault, and with the Beinn Bhreac fold (Fleuty 1974), which lies to the NE. It has a NE-SW trending axial trace in its type area, and a moderate, reclined SE plunge, its limbs appear to be parallel to the margins of the pluton, especially at the SW end of the igneous complex. Ramsay presented evidence that these F_3 folds were later than early ductile movements on the Strathconan Fault, and hence were intimately associated with the fault. This hypothesis was supported by Sutton and Watson (1958) who coined the phrase 'Strathconan disturbance' to describe the complexity of the geology in this area. Fleuty, in discussion of Ramsay (1958) postulated that the Strathconan Fault may overlie a deep-seated line or belt of weakness in the basement, above which there are several "twists" suggestive of sinistral shear. A similar zone of complexity is associated with the Great Glen Fault (Watson 1984). Nowadays, the F_3 structures are correlated with the folds of the North Highland Steep Belt (Leedal 1952; Barr *et al.* 1986), which are known to fold the Sgurr Beag and other ductile thrusts in the Moine (Barr *et al.* 1986). They also deform the Glen Dessary syenite (c. 456 Ma.; Roberts *et al.* 1984). These Steep Belt fold structures are then cross-cut by a series of pegmatites

which have been dated at c. 430 Ma. (van Breeman *et al.* 1979b). If this is the case and the Strathconan Fault did not start to move until c. 430Ma (Watson 1984), then it is unlikely that these folds were produced by sinistral movements along the fault, but it does not exclude the possibility that their geometries may have been modified in response to sinistral shearing. This idea of a deep-seated basement structure along the trace of the Strathconan Fault would certainly be very conducive to magma transport if not magma production (Hutton & Reavy 1992) and may, therefore, be extremely influential in the location of plutonism in the area of the fault. Furthermore, the emplacement of the c.425 Ma Ratagain complex in association with sinistral movements along the Strathconan Fault using the pre-existing anisotropy produced by F_3 folding, seems to fit the regional deformation chronology, outlined above, very well.

Figure 5.13 The geometry of the Beinn a' Chaonnich antiform (after Ramsay 1958). Compare with Fig. 5.12 illustrating the pluton geometry.



The present author tentatively suggests that the form of the pluton is strongly influenced by the geometry of the Beinn a' Chaonnich fold (as illustrated in Figs. 5.11 & 5.13). This relationship is primarily seen by the parallelism of the pluton margins and the foliation in the country rocks. In addition, however, the pluton's flat roof could be explained in terms of the geometry of the fold structure, if the magma passively followed the country rock anisotropy during localised transtension along the Strathconan Fault. The geometry of the F_3 structures can be inferred in three dimensions by careful examination of the maps of the area, which reveals a zone of NE-SW strikes, dipping shallowly to moderately NE along the SW shore of Loch Duich. Minor F_3 folds are very common in this area (Sutton & Watson 1958), where they nearly always plunge NE, as opposed to SE in the SW sector of the pluton, so the fold is curvilinear (see Fig. 5.15). An interesting point to note is that the contacts of the pluton in the NE appear to parallel the NE plunge of the F_3 folds in this area, whereas, in the SW the contacts plunge SE, again parallel with the plunge of the F_3 structures in this area (see Figs. 5.10 & 5.13), so the curvilinearity of the fold may be directly reflected in the change of strike and dip of the pluton contacts. This is a tentative proposal, since the structure of the country rocks in NE of the pluton is extremely complex due to the presence of significant F_2 folds, such as the Letterfern antiform, which may also influence the form of the pluton in this NE sector.

5.6 Conclusions

- (1) Regional Caledonian F_3 fold structures in the Moine and Lewisian country rocks, especially the Beinn a' Chaonnich antiform may have been extremely influential in determining the geometry of the Ratagain pluton.
- (2) the Ratagain Igneous Complex was emplaced at C. 425 Ma. and was deformed during emplacement by sinistral movements along the Strathconan Fault and associated WNW-ESE trending synthetic splays, producing well-developed magmatic state deformation fabrics within the pluton.
- (3) It is proposed that space was created for pluton emplacement due to differential rates of movement along the strike of the Strathconan Fault, in response to rheological weakening of parts of the fault due to the presence of magma at depth. At the present level of exposure, there is no evidence to suggest involvement of the Moine Thrust during emplacement.
- (4) A hiatus in sinistral movement along the fault occurred after emplacement, noted by the absence of solid state deformation fabrics within the pluton. Sinistral movement resumed at about 400 Ma., resulting in the development of cataclastic and breccia zones within and outside the pluton.
- (5) This study illustrates the importance of interactions between active tectonism and pre-existing structures within metamorphic basement terrains of this kind. In particular, these interactions may directly control the siting and geometry of late orogenic intrusions at various scales.

Chapter 6

THE LOCH LOYAL SYENITE COMPLEX

Introduction

The Loch Loyal Syenite Complex forms an impressive area of mountainous terrain, approximately 9km south of Tongue, northern Sutherland (Fig. 6.1). It was remapped, as part of the NERC/BGS Academic Mapping Programme (Grant F60/G2/36), sheet 114 East, by R.E. Holdsworth, R.A. Strachan, G.I. Alsop and the present author. The present author was responsible for mapping the Cnoc nan Cuilean body, the eastern, northwestern and western areas of the Ben Loyal body, reconnaissance mapping of the Ben Stumanadh body, and examination of the deformation textures of all three bodies in thin section.

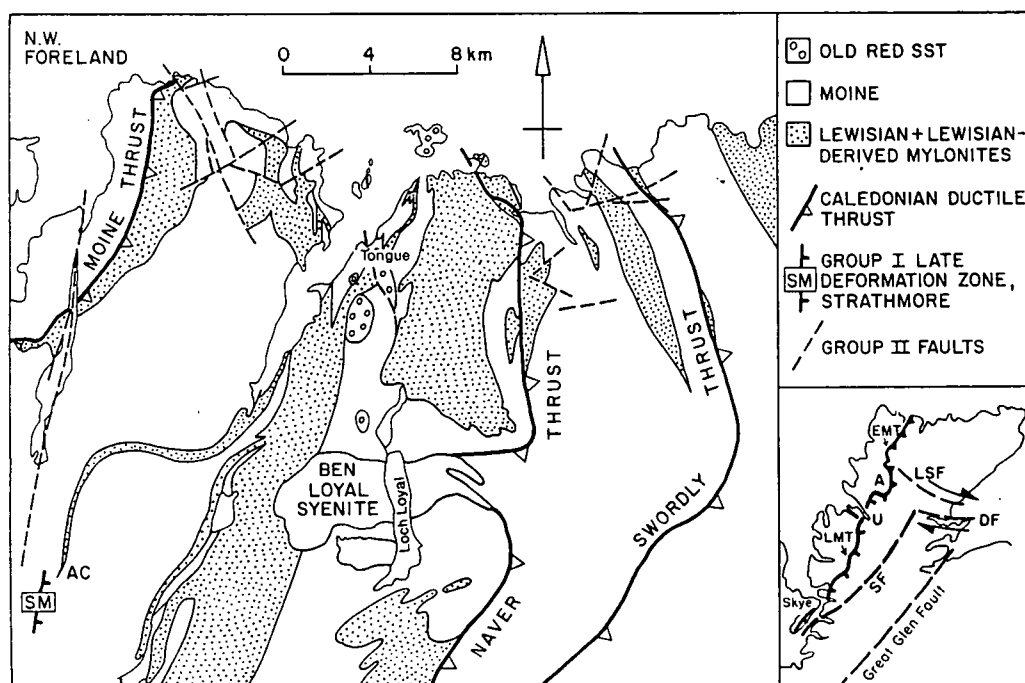


Figure 6.1 Regional geological map of the N. Sutherland area, including the area covered by NERC/BGS Academic mapping contract (Grant F60/G2/36), which includes the Loch Loyal Syenite Complex (after Holdsworth 1989).

6.1 Regional Setting

The Loch Loyal Syenite Complex is composed of three distinct bodies, the Ben Loyal intrusion, the Ben Stumanadh intrusion and the Cnoc nan Cuilean body (Fig. 6.2 & Map 4), each of which have slightly different compositions, but cover a combined area of approximately 24km². The Ben Loyal Intrusion is the biggest of the three (16km²), making it the largest alkaline intrusion in Britain. The syenites lie 12km east of the present day trace of the Moine thrust (Fig. 6.1) and their petrographical similarity to the alkaline igneous complexes in Assynt (Loch Ailsh, Loch Borrolan) has been noted by several previous authors (e.g. Read 1931, 1934; Phemister 1936; 1948; King 1942); they are chemically distinct, however (Parsons 1972; Robertson & Parsons 1974). They were emplaced into Moine and Lewisian rocks, that had previously been metamorphosed to amphibolite facies and deformed during Caledonian NW-directed shearing (Read 1931, 1934; Fig. 6.2 & Map 4). In addition to the main plutonic units there are numerous dykes, veins and sheets of syenite in the surrounding country rocks, occurring up to 5km away, e.g. the region N of Lochan na Cuilce, 4km NNW of Ben Heil (Holdsworth 1987). Radiometric ages have been given by several investigators; Brown and Miller (1965), 378±17Ma on hornblende; Brown, Miller and Grasty (1968), 404±10 & 402±9Ma on pyroxenes; Halliday *et al.* (1984), 426±3Ma from U-Pb on zircon from the Cnoc nan Cuilean body. For the purposes of the present study, the latter age is taken as being the emplacement age of the entire complex.

Subsequent to intrusion, the igneous complex was uplifted and eroded, together with the country rocks, so that syenite boulders form a major clast type found in nearby conglomerate breccia deposits of Old Red Sandstone age (Map 4; O'Reilly 1971; Blackbourn 1981). In addition, remapping of the area resulted in the discovery of a small ORS deposit resting directly on top of the Ben Stumanadh syenite, near the summit of Ben Stumanadh.

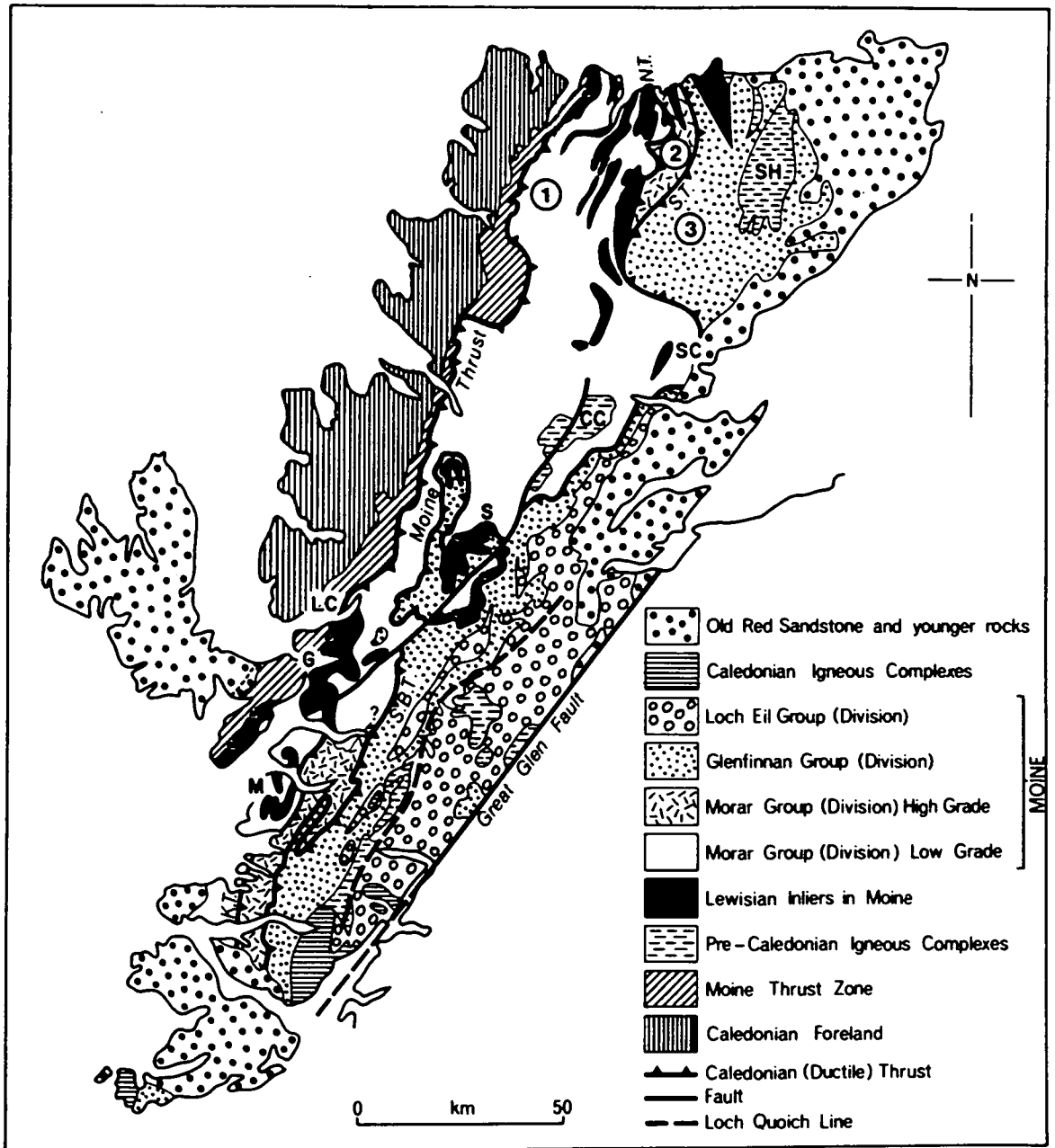


Figure 6.2 Stratigraphical and structural units in NW Scotland. Map also shows the position of the Moine Thrust, the Naver Thrust (NT) and the Swordly Thrust (ST).

6.2 Country rock stratigraphy and structure in the Loch Loyal area

The regional geology of the Moine rocks of the Northern Highlands is outlined in Chapter 5. The Loch Loyal syenites were emplaced into Moine rocks including numerous Lewisian inliers tectonically emplaced both as slices along ductile thrusts and as the cores of major folds. Earlier workers favoured a Precambrian age of emplacement (Moorhouse 1977; Mendum 1979; Moorhouse & Moorhouse 1979), however, more recent studies have suggested that a predominantly Caledonian age of emplacement is more probable on the basis of structural kinematics and isotopic age data (Barr *et al.* 1986; Strachan & Holdsworth 1988; Holdsworth 1988, 1989; Moorhouse *et al.* 1988). The Loch Loyal intrusions cross-cut the upper parts of the Caledonian Moine Nappe and the lower part of the Naver Nappe in Sutherland (Fig. 6.1 & 6.2).

The regional geology of the rocks of the Moine Nappe have been discussed recently by many authors including Barr *et al.* (1986), Moorhouse *et al.* (1988), Strachan & Holdsworth (1988) and Holdsworth *et al.* (1993). Holdsworth (1987, 1989, 1990) and the BGS mapping contract, out of which this study arose, have considered the detailed structure of the area around the Kyle of Tongue and the salient features are discussed below in relation to the main phase of Caledonian ductile thrusting.

Pre-thrusting deformation

Pre-thrusting, D_1 structures in the Moine nappe of Sutherland are dominated by a well developed bedding-parallel foliation, S_1 . There is some evidence of heterogeneous strain during this early deformation since S_1 is most intense in areas closest to original stratigraphic Moine - Lewisian contacts (e.g. the Strathan conglomerate; Mendum 1976). In many areas, D_1 strains are interpreted as being low, since sedimentary structures are commonly preserved within the Moine succession (Holdsworth 1987, 1989). The Moine pelites and the Ben Hope Sill amphibolites, a set of pre-tectonic (Holdsworth 1987, 1989) garnet amphibolites (Moorhouse & Moorhouse 1979; Floyd & Winchester 1983), carry early, well defined S_1 fabrics as their primary metamorphic fabric, which are consistently deformed by D_2 structures (Holdsworth 1987, 1989). D_1 folds are only small-scale, minor features, which, in the Moine

metasediments, are difficult to confidently distinguish from syn-sedimentary slumps (Holdsworth 1987, 1989). Furthermore, in Lewisian inliers, it is difficult to assign relatively rare examples of pre- D_2 folds to D_1 since they could be equally representative of pre-Moinian basement structures. Moine psammites also often have a N-to NNE-trending mineral lineation, which is believed to be D_1 in age since it is refolded by F_2 folds (Holdsworth 1987, 1989).

Ductile thrusts and associated fold structures

Main Phase

Fabrics and folds directly related to the main phase of Caledonian ductile thrusting usually form the second phase of structures recognised locally in exposures, D_2 . This deformation phase is associated with ubiquitous development of a gently ESE-dipping schistosity. In micaceous and hornblendic rock types, this fabric forms a tight crenulation cleavage that often becomes transposed in areas of intense strain. The schistosity intensifies into broad zones of high strain, up to 80m thick, without significant reorientation, in association with ductile thrusts, resulting in the development of belts of platy mylonites (Holdsworth & Grant 1990). Lewisian inliers are often carried in the hangingwalls of ductile thrusts as deformed thrust slices. An ESE- to SE-plunging mineral lineation is developed throughout the Moine Nappe in association with D_2 deformation, being most strongly developed in the mylonite zones. Occasional relict *S-C* fabrics are preserved locally within the mylonite zones, shear sense indicators (mica fish, shear bands, etc.) suggesting overthrusting towards the WNW-NW, parallel to the observed mineral lineation (Holdsworth 1989). F_2 folds dominate the structure of the Moine rocks at all scales, the S_2 schistosity being axial planar to these folds. Detailed structural analyses (Holdsworth 1987, 1988, 1989) indicate that the folds formed as WNW- to NW-overtaken buckles, with hinges lying at high angles to the direction of thrusting. Shearing subsequent to the development of the folds in association with movements on ductile thrusts resulted in rotation of fold axes towards the tectonic transport direction (c.f. Escher & Watterson 1974), so that the majority of F_2 hinges now plunge shallowly to the ESE, sub-parallel to the Caledonian mineral lineation. As a consequence, these folds also display tight sheath fold geometries on all scales (Holdsworth 1990), as exhibited by the preservation of 'eye structures' (Cobbold

& Quinquis 1980). Many other Lewisian inliers in Sutherland lie at the cores of some larger Caledonian anticlinal sheath folds (e.g. see Holdsworth 1988).

Secondary Phases

F_3 folds are of lesser importance regionally than F_2 structures. However, in many areas in Sutherland, especially in the region around the Kyle of Tongue, a series of F_3 folds are seen to deform the main phase Caledonian folds and their associated linear and planar fabrics (Holdsworth 1990). These local F_3 structures range in scale from millimetres to hundreds of metres, but are mostly restricted to the ductile thrust high strain zones, for example, the ductile thrust imbricate in the Talmine - Ben Hutig area, the immediate hangingwall of the Ben Hope Thrust and in the high strain zone adjacent to the Ben Loyal syenite.

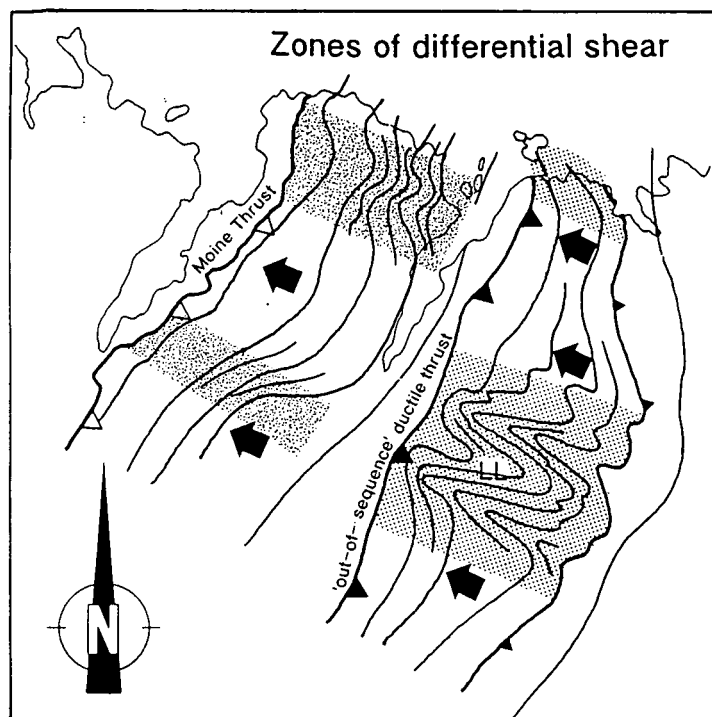
Holdsworth (1990) defines two types of minor F_3 fold geometry that are observed in the Kyle of Tongue area both of which he relates to the development of localised flow perturbations during ductile shearing (Fig. 6.3).

(a) *sheath fold types* which are highly curvilinear, having up to 160° hinge curvature over distances of a few metres about an ESE-plunging axis, and are close to isoclinal. Where they are preserved in their original NNE-SSW orientation, they are always overturned towards the WNW and refolding of main phase folds and mineral lineation is well developed. These structures are thought to form due to changes in the rate of ductile flow parallel to tectonic transport (Holdsworth 1990).

(b) *asymmetric types* that are gentle-to-tight, plunge mostly eastwards and show very little obvious hinge curvature. Fold axes lie either at low angles or parallel to the main phase mineral lineation, although refolding of the mineral lineation may be obscured by the development of a strong rodding fabric parallel to local F_3 axes (c.f. Wilson 1953; Holdsworth 1987). Although these asymmetric types have hinges that lie close to the direction of tectonic transport, the degree of fold axis rotation due to shearing is unlikely to be substantial because interlimb angles are usually large ($\geq 60^\circ$) and strain is low. Holdsworth (1990) relates these folds to changes in the rate of flow in a direction normal to tectonic transport, i.e. differential shearing during ductile thrusting.

Most large-scale F_3 folds are asymmetric types, usually with large interlimb angles ($\geq 90^\circ$). These tend to affect areas of uniform rock units outside zones of intense deformation, and where observed, they produce marked changes in the strike of the regional foliation (e.g. Fig. 6.3). As a result, they are referred to here as *cross-folds*. The formation of these zones of cross-folding is believed to be associated with larger-scale differential shearing during ductile thrusting (Holdsworth 1990). Although these structures post-date the main F_2 fold phase, they are believed to be late D_2 in age. Large-scale cross-folds further to the west root into the Moine Thrust Mylonite Zone (Alsop & Holdsworth 1993), which is believed to be pre-430Ma, based on cross-cutting relationships with the Loch Borrolan Syenite (Coward 1983, 1984).

Figure 6.3 Sketch diagram showing zones of anomalous strikes in Sutherland. The anomalous strike is due to the presence of cross folds.



The Loch Loyal Syenite Complex is spatially associated with a zone of cross-folds, which are believed to relate to a high strain zone running along the NW side of the intrusion, running immediately above one of the numerous Lewisian inliers in the region. This zone of high strain is believed to be contemporaneous with the ductile thrusting responsible for the production of the Moine Thrust Mylonites.

It was formed at amphibolite facies, and may represent a significant out-of-sequence structure (Holdsworth pers. comm.). Although the age of cross-fold development pre-dates the intrusion of the syenite, effectively excluding active coincidence of cross-fold development and magmatism, the present investigator believes that the change in strike of the regional foliation to an anomalous NW-SE orientation as a result of cross-fold formation may play an important role in facilitating the emplacement of the igneous complex (see section 6.5.2 for further details).

Post-thrusting deformation

Relatively little attention has been paid to late, brittle deformation features which are frequently observed in Sutherland. Two distinct groups of post-thrusting structures are recognised within the Moine nappe (O'Reilly 1971; Holdsworth 1987, 1989):

- (a) *Group I Structures*: a set of brittle folds and low angle faults with various orientations;
- (b) *Group II Structures*: these post-date Group I structures. They are a set of high angle, predominantly extensional faults, which are especially common in the region of the north Sutherland coast, partly due to better exposure here.

Group I Structures

These have been loosely described as 'D₄' (e.g. Soper 1971), although O'Reilly (1971) identified two sets of predominant structures belonging to this subdivision of post-thrusting deformation, D₄ and D₅, which can be distinguished on the basis of orientation. Both sets of group I folds display similar complex kink, conjugate and box-fold geometries (Holdsworth 1987, 1989) and they are all closely associated with faults, which may take the form of small-scale fractures or larger scale detachment zones.

Patterns of detachment ramp and associated fold orientation suggest an ESE-transport direction, that is in the opposite direction to the thrust transport direction. These observations lead Holdsworth (1987, 1989) to suggest that the ESE displacements responsible for the formation of Group I structures occurred due to gravitationally-induced back-collapse and thinning of the Moine Thrust sheet at a late stage of Caledonian orogenesis. This is a particularly important process, which may be linked to the emplacement of alkaline magmas in this area (see sub-section 6.5.2).

Group II structures

These are vertical or steeply-dipping faults of various orientations, ranging from WNW-ESE, NW-SE, NNW-SSE, to NE-SW. The variably orientated fault segments form a linked-system and display predominantly extensional displacements of a few metres or more, based on recorded offsets of basement units (Holdsworth 1987, 1989).

The general consensus is that the most steeply dipping fractures recognised in Sutherland are Devonian extensional faults, which are a weak onland manifestation of the crustal stretching responsible for the formation of the offshore West Orkney Basin (Coward & Enfield 1987; Enfield & Coward 1987). O'Reilly (1983) and Watson (1984) suggested that the present-day onshore land surface corresponds closely with that of the Devonian, which would imply that these Group II faults formed very close to the surface. This is supported by the occurrence of incohesive gouge and breccia fault rocks where fault zones are exposed (Holdsworth 1987, 1989). This style of deformation is consistent with near-surface fault movements (Sibson 1977; Hancock *et al.* 1987). It is believed that the NE-SW trending late fault cutting the Loch Loyal complex (see Map 4) is related to these structures (see subsection 6.4.3 for discussion of this).

6.3 Petrography and geochemistry of the Loch Loyal syenites

Various aspects of the petrography and geochemistry of the syenite complex have been studied in detail during the past century. Its petrography was compared with that of the Loch Ailsh syenite by Read (1931). King (1942) carried out a detailed petrographical and geochemical study of the Cnoc nan Cuilean body (Fig. 6.?). Parsons (1972) compared the normative mineralogy of the Loch Loyal syenites to those in Assynt, and showed real chemical differences between the two areas. Parsons and Boyd (1971) briefly described the feldspar variations within the pluton. Finally, Robertson and Parsons (1974) gave a relatively detailed account of the structure and petrology of the Ben Loyal syenite, with brief discussion of the petrographical and chemical variation exhibited by the Ben Stumanadh and Cnoc nan Cuilean syenites.

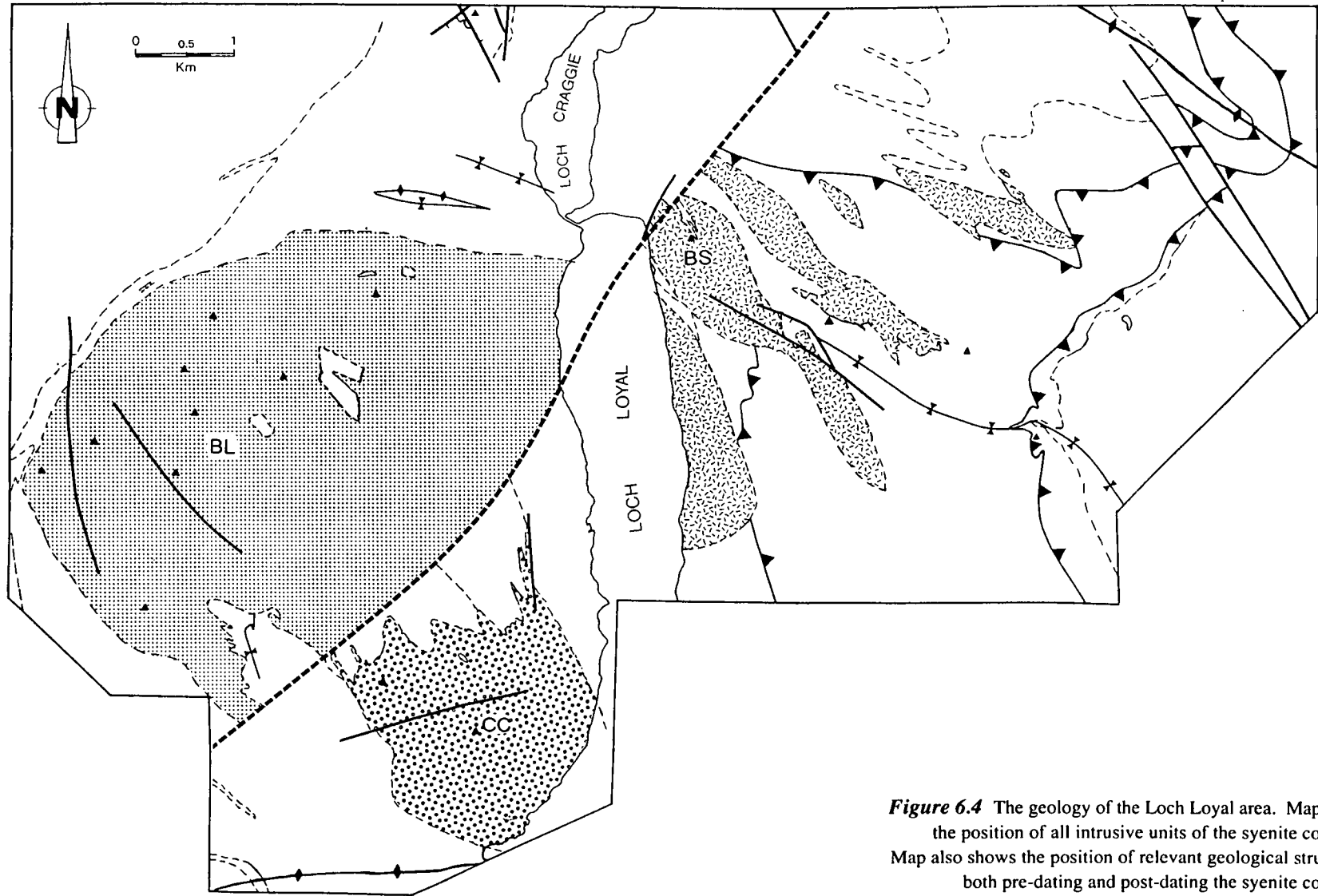


Figure 6.4 The geology of the Loch Loyal area. Map shows the position of all intrusive units of the syenite complex. Map also shows the position of relevant geological structures, both pre-dating and post-dating the syenite complex.

6.3.1 Petrography

The petrography of each of the syenite units varies slightly, but all show alkaline affinities. No detailed petrographic studies were carried out during the present work. However, general observations made during examination of the fabrics in the syenites (detailed in section 6.4) largely confirm the observations made by Parsons and co-workers. These data are summarised here, with additional new observations where appropriate.

Ben Loyal

The Ben Loyal body comprises two quartz syenites, in a roughly semi-circular shaped intrusion; an outer well foliated unit and an inner less well foliated unit (see Map 4 & Fig. 6.4). The contact between the two units is gradational both in the field and in thin section, with the inner unit representing between $\frac{1}{3}$ to $\frac{1}{2}$ of the intrusion. The details of the foliation and its possible origin will be discussed further in a later section of this chapter.

Both units are relatively pale in colour, their mineralogy being dominated by feldspar. The most abundant mafic mineral in both the units is aegirine-augite, with the exception of some marginal phases where amphibole is more abundant, perhaps due to an increase in water vapour pressure in this facies, as suggested by Robertson and Parsons (1974, p.139). Amphiboles occur as irregular prisms 1-3mm long and include hornblende and an alkali amphibole which is a member of the eckermannite-arfvedsonite series and shows some variation in extinction angle between rocks (Robertson & Parsons 1974). In some of the core syenites colour zoning of mafic minerals occurs, from pale centres to green margins. This phenomenon is more common in the Ben Stumanadh and Cnoc nan Cuilean syenites.

Biotite occurs only occasionally in the outer syenite unit, often as pale deformed flakes, or showing alteration to chlorite (Robertson & Parsons 1974). The most abundant accessory mineral is sphene, together with apatite, allanite and ore. In both syenite units of the Ben Loyal body, there is a tendency for mafic and accessory minerals to occur together in clots of a number of grains of several mineral species (Robertson & Parsons *op cit.*). Von Knorring and Dearnley (1959) reported the presence

of a canary yellow, kaolin-like, powdery rare earth mineral, thought to be a hydrated form of monazite. It occurs, often intergrown with apatite, in some, but not all, miarolitic cavities, which are more-or-less restricted to the marginal syenites of the Ben Loyal body, often along foliation planes. The present study shows that vugs and miarolitic cavities are much more common in the Ben Stumanadh body, where they are often lined with carbonate and subordinate amounts of the rare earth mineral described above. Von Knorring and Dearnley (1959) suggested that the mode of occurrence and mineral association suggest a low temperature, hydrothermal origin for this rare earth mineral.

Quartz is fairly constant in amount throughout both intrusive units. However, in the outer unit it is coarse grained and often visible in hand specimen. In thin section, these grains are often anhedral and show pronounced undulose extinction, which may be a product of deformation. It is important to remember that the formation of such textures is enhanced by elevated water vapour pressures. In contrast, in the core rocks it occurs as groups of small, rounded grains in association with albite between large, perthitic alkali feldspars; undulose extinction is less common in the quartz of the inner intrusive unit. In addition to quartz crystals, thin veins of quartz are not uncommon, especially in the outer unit in the area around Lettermore quarry and Ben Heil (Robertson & Parsons 1974).

According to Parsons and Boyd 1971 and Robertson and Parsons 1974, the most important difference between the inner and outer syenite units of the Ben Loyal body is that the outer body is a two feldspar, subsolvus assemblage rock, whereas the inner unit bears only one feldspar in appreciable quantities, and appears to be hypersolvus. Table 6.1 outlines the modes of representative Ben Loyal syenites, as determined by Robertson and Parsons (1974).

Table 6.1 Modal analyses of representative Ben Loyal syenites (Vol %).

	CORE SYENITE	OUTER SYENITE
Quartz	8.5	10.2
Alkali feldspar (including perthitic albite)	80.8	55.1
Separate Plagioclase	4.0	23.8
Pyroxene	4.3	5.6
Amphibole	0.0	3.7
Accessories (see text)	2.4	1.6

The outer unit contains abundant discrete plagioclase together with alkali feldspar, while the core syenites contain a single, perthitic feldspar, with only minor interstitial albite (Robertson & Parsons 1974). These authors document convincing textural evidence that the two-feldspar rock developed, at least in part, by unmixing of initially perthitic alkali feldspar in the manner envisaged by Tuttle (1952). C.I.P.W. norms of the Ben Loyal syenites show little variation in the bulk feldspar composition and no systematic difference between the inner and outer syenite units, effectively eliminating the possibility that the appearance of two feldspars is related to bulk anorthite content or to increasing SiO_2 . The appearance of discrete plagioclase is balanced by a decrease in the amount of albite in perthitic intergrowth. Phemister (1926) and Parsons (1965) have described the development of two feldspar assemblages in the Loch Ailsh syenites, which have been subjected to post-intrusive shearing and, in places mylonitisation. The occurrence of a two feldspar assemblage only in the foliated unit of the Ben Loyal intrusion must surely pose the question is subsolidus unmixing of feldspars a deformation enhanced process? Robertson and Parsons (1974) contend that in the Ben Loyal intrusion the process is related to deformation produced by the intruding magma. This is a hypothesis supported by the present author, but amended to include the effects of regional tectonics, which were, at least partly responsible for creating the space for intrusion of the magma and the fabrics subsequently developed within the intrusion (see section 6.5).

Robertson & Parsons (1971) considered that the slightly finer grained nature of the outer syenites could be attributed to the change from a coarse, single feldspar rock in the core to a rock containing two feldspars. The current work suggests that the coarser grain size of the inner unit could also be attributed to slightly slower cooling rates in the inner unit due to its slightly later intrusion into the centre of the partially molten outer unit, as opposed to intrusion into relatively cold country rocks.

During the present study, it has been observed that the marginal phase of the Ben Loyal body contains more enclaves of basic material than the syenites towards the core of the body. These mafic xenoliths resemble those observed throughout the Cnoc nan Cuilean body (see later in this section).

Ben Stumanadh

Detailed petrological and chemical studies on the Ben Stumanadh body are lacking. Robertson and Parsons (1974) made some basic petrological observations in order to compare this with the Ben Loyal intrusion, and the current mapping project has made some further observations. In the field and hand specimen, the Ben Stumanadh syenite is a dark brownish- to pinkish-red colour. It is emplaced on all scales as a series of NW-SE trending, en echelon, variably coalesced dykes, which are usually sub-parallel to the foliation in the Moine country rocks (Fig. 6.4 & Map 4). The dykes carry a relatively well developed foliation which is sub-parallel to the dyke walls in larger sheets, but occasionally oblique in narrower dykes; the significance of these fabrics and their geometry will be discussed fully in section 6.4.3.

Petrologically, the Ben Stumanadh rocks exhibit similar, although not identical characteristics to the Ben Loyal rocks. Robertson and Parsons (1974) reported the presence of extensive marginal zones to many sheets, which they interpreted as the result of hybridisation, since feldspathised schists grade into normal syenite. These authors claimed that the hybrid rocks contained two feldspars, coarse non-perthitic microcline co-existing with large, finely twinned plagioclase, whilst rocks from the interior of the main sheet are very similar to the inner unit of the Ben Loyal syenite, since they have only variable marginal albite between regularly perthitic alkali feldspar, which may show oscillatory zoning. However, the ten specimens of Ben Stumanadh syenite examined during this project indicate that the occurrence of two feldspars, perthitic alkali feldspar and finely twinned plagioclase, is not uncommon; microcline is rare. There is, however, a variation in the amount of well developed, discrete crystals of plagioclase from place to place. It is suggested here that the Ben Stumanadh syenites, in terms of feldspar composition, bear a stronger resemblance to the outer, marginal, unit of the Ben Loyal intrusion, than the inner unit.

Quartz is fairly constant in proportion, as in the Ben Loyal syenite. However, in contrast to the Ben Loyal intrusion, it is sometimes closely associated with feldspar in a patchily developed graphic intergrowth. The mafic mineralogy is different from Ben Loyal, since amphibole is always found associated with pyroxene, again this is possibly due to elevated water vapour pressures in this unit. As in Ben Loyal, amphibole minerals are hornblende and bright green alkali amphibole.

The Ben Stumanadh syenite has a very pronounced vuggy texture. In contrast to Ben Loyal, the vugs are believed to have developed at more than one time. An early set are believed to be analogous to the miarolitic cavities seen in Ben Loyal. These cavities contain quartz, apatite, zeolites and occasionally, the yellow rare earth mineral similar to that described by von Knorring and Dearnley (1959). A second set of vugs are believed to result from the retrogression of mafic minerals due to percolation of fluids in relation to movements on the late NE-SW trending fault (see Fig. 6.4 & Map 4). These are particularly prolific close to the shores of Loch Loyal, in close proximity to the late NE-SW striking major fault, where they are filled with clay-like minerals and carbonates, similar to those found as fault gouge and fault precipitated material respectively (see Plate 6.1). However, there are no cross-cutting relationships of the two proposed sets of vugs, which would indicate without doubt that these are two distinct generations of cavities.

Cnoc nan Cuilean

According to Robertson & Parsons (1974), this body has a 'stock-like' form and appears to be separated from the Ben Loyal intrusion; King (1942) had previously contended that the two were connected. The intrusion has relatively basic margins which grade into a more homogeneous feldspathic central portion, although overall this syenite body has a more heterogeneous composition than either the Ben Loyal or Ben Stumanadh intrusions. Robertson and Parsons (1974) also showed that the petrology and composition of the Cnoc nan Cuilean body differs slightly from the Ben Loyal intrusion. King (1942) attributed such differences in composition to the belief that the Cnoc nan Cuilean body had a significant metasomatic component. He reported that the contacts with the country rock were often gradational, due to a transition from metasomatically altered siliceous Moine psammities, into a marginal, basic syenite. He reported that this gradation was particularly well developed in a line of crags extending from the falls in Allt Torr an Tairbh, above Loch Loyal Lodge, and around the upper slopes of Lettermore and Meall Eudainn (see Map 4). King believed that the marginal syenites were a product of hybridisation of highly modified, basic metasomatic rocks from the Moine and a feldspathic syenite. Remapping of the country rocks in this region demonstrates, however, that the majority of so-called hybridised country rocks (King

1942) are banded hornblende-pyroxene gneisses and amphibolites of typical Lewisian aspect (Holdsworth pers. comm.). Thus, although very localised alkali amphibole and pyroxene growth is observed in some contact rocks (e.g. NC 5985 4547), the degree of metasomatic alteration is far less extensive than that proposed by King (1942). In addition, Robertson and Parsons (1974) analyses some relatively feldspathic, xenolith-free Cnoc nan Cuilean syenite, which seemed to suggest a primary magmatic difference in the Cnoc nan Cuilean syenite and the Ben Loyal syenite.

The marginal syenites of Cnoc nan Cuilean consist of a two feldspar, subsolvus assemblage, rather like the two feldspar assemblage in the outer unit of the Ben Loyal intrusion. The mafic mineralogy is totally dominated by aegirine-augite. Cnoc nan Cuilean appears to be devoid of amphibole in the marginal facies, in contrast to Ben Loyal and Ben Stumanadh. The main syenite is a one feldspar, hypersolvus syenite, with aegirine augite as the dominant mafic phase. Accessories in both the marginal and main facies are similar to the Ben Loyal intrusion. The Cnoc nan Cuilean syenite does not appear to contain as many vugs/miarolitic cavities as the other intrusive units. Xenoliths of country rock are relatively common throughout the Cnoc nan Cuilean body and are not only confined to the marginal basic syenite, as reported by King (1942).

6.3.2 Major element chemistry of the Loch Loyal syenites

The only detailed study of the chemistry of the three syenite intrusions was carried out by Robertson and Parsons (1974). The results of their study are briefly summarised in the following section. The Loch Loyal syenites are all of alkaline affinities. Robertson and Parsons (1974) reported the results of XRF analyses and CIPW norms of 29 'normal' Loch Loyal syenites, chosen to be free of xenolithic material, results are summarised in Fig. 6.5. Marginal variants and aplite were analysed separately, and salient features are also summarised in Fig. 6.6, together with results from the inner and outer Ben Loyal units.

Figure 6.5, according to Robertson and Parsons (1974), illustrates that the Loch Loyal syenites, when considered in terms of *ab-or-Q*, plot close to a thermal valley in the Ab-Or-Qz system determined by Tuttle and Bowen (1958). When normative *an* is considered they lie within or very close to the alkali

feldspar volume in Ab-Or-Qz at $P_{H_2O} = 1\text{Kb}$ (c.f. Carmichael 1962; James & Hamilton 1969; Parsons 1972 for assumptions used in construction). In the opinion of Robertson and Parsons (1974) the restriction of all rocks of the Loch Loyal suite to this phase volume suggests that the dominant control on rock composition has been magmatic, with no extensive redistribution of alkalis in a volatile phase.

Figure 6.5 Normative $Q\text{-}ab\text{-}or\text{-}an$ for analysed Loch Loyal syenites plotted on triangular diagrams showing field boundaries (solid lines) and unique fractionation curve (broken line) in the synthetic Ab-Or-Qz and Ab-Or-An systems, at $P_{H_2O} = 1\text{Kb}$. Rocks with zero an have been omitted from the lower diagram (after Robertson & Parsons 1974).

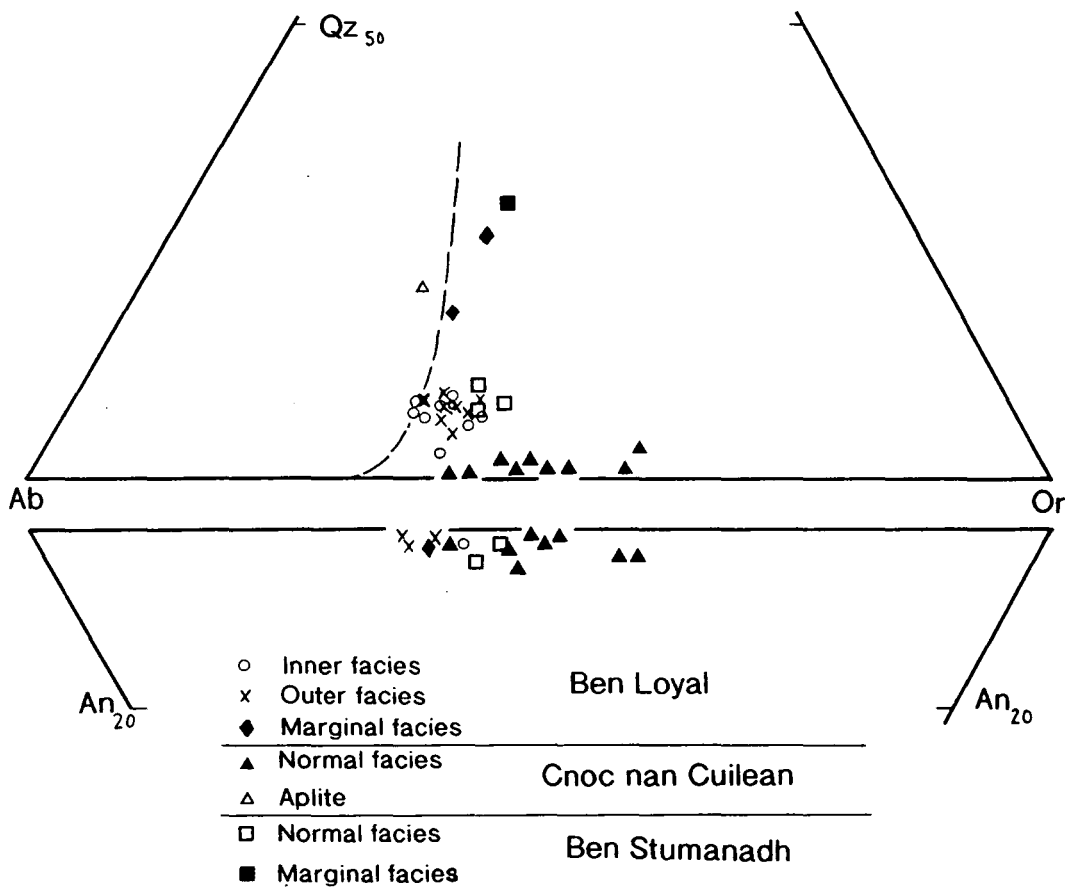
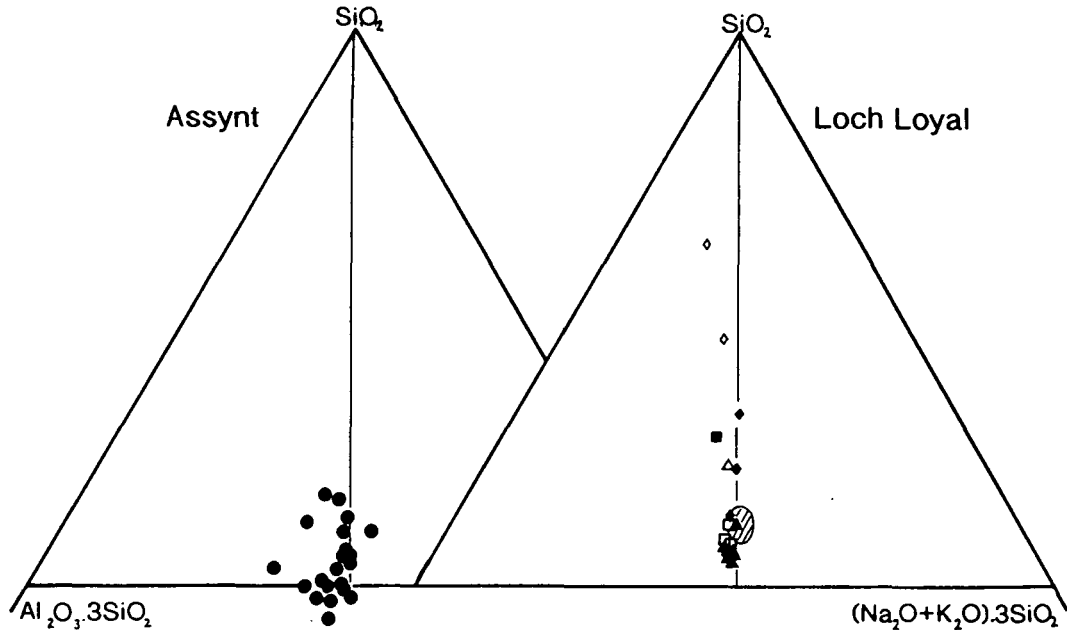


Figure 6.6 Molecular $\text{SiO}_2\text{-Al}_2\text{O}_3\text{-(Na}_2\text{O + K}_2\text{O)}$ of NW Highlands leucosyenites plotted on the diagram of Bailey & MacDonald (1969). The symbols used are those of Fig. 6.5. In addition open diamonds are psammites, solid circles are rocks from the Loch Ailsh and Loch Borallan complexes. All Ben Loyal rocks lie in the shaded area (after Robertson & Parsons 1974).



Robertson & Parsons (1974) suggested that samples from the Ben Stumanadh syenite lay on the potassic side of Ben Loyal group rocks, suggesting that Ben Loyal and Ben Stumanadh were separate bodies. However, the present author believes that the difference is minor and, if real, could be due to limited, although not extensive redistribution of alkalis in a volatile phase. This is supported by the fact that the laminated outer unit of the Ben Loyal syenite plots around the same area as the 'normal' (*sensu* Robertson & Parsons 1974) Ben Stumanadh rocks, and the 'marginal' syenites of the Ben Loyal syenite plot in roughly the same area as the 'marginal' facies of the Ben Stumanadh syenite. Indeed, Robertson and Parsons (1974) proposed the idea of a water gradient from core to margin within the Ben Loyal intrusion, suggesting that this was one of the conditions that contributed to the unmixing of the perthitic feldspar, to a two feldspar, subsolvus assemblage. The current author believes that such a water gradient is likely to have existed, not only towards the 'lateral' margins, but also upwards to the top of the magma body, facilitating the same reactions above the present level of exposure of the Ben Loyal intrusion. It is suggested in section 6.5 that the Ben Stumanadh syenite forms part of the same igneous body as Ben Loyal, representing a higher structural level exposed in its present position due to later movements along the NE-SW Loch Loyal Fault, which has a significant component of downthrow to the southeast. This

model would account for the similarities, both chemically and petrologically between the Ben Loyal and Ben Stumanadh syenites, that is, a tendency towards more potassic chemistry; the same feldspar assemblage; the occurrence of amphibole as well as pyroxene; the development of graphic intergrowth of quartz and feldspar and the presence of vuggy texture. Textural evidence which further corroborates this hypothesis will be presented in section 6.4.

Cnoc nan Cuilean is seen to be more mafic in terms of its chemistry than the other Loch Loyal syenites, confirming that its more basic appearance is due, at least in part, to its initial composition, and not simply as a result of hybridisation of country rock material, as suggested by King (1942). Robertson and Parsons (1974) suggested that this was further indication that Cnoc nan Cuilean is a separate intrusive unit, and further, that it is relatively less fractionated and a drier intrusion; this is supported by its mineralogy and petrology (notably the absence of amphibole). Consistently higher radiometric anomalies over Cnoc nan Cuilean, demonstrated by Ball and Plant (Gallagher *et al.* 1971), also suggest that this body is chemically distinct. The present author supports these ideas, since field observations indicate that it has no connection with Ben Loyal, being separated from the latter by the late NE-SW trending fault, but with no evidence of any connection between the two before the occurrence of the late fault. The total absence of amphibole suggests that there was rather less water available in this intrusion than elsewhere in Loch Loyal. The localised occurrence of mafic syenite as enclaves within both the Ben Loyal and Ben Stumanadh syenites seems to suggest that the Cnoc nan Cuilean body is the earliest of the Loch Loyal syenites.

6.4 Intrusion geometries and deformation features in the syenite complex

6.4.1 Contact relationships

Some basic observations were made about the contact relationships between the country rocks and the syenite bodies in the field.

- (a) The syenites cross-cut *all* the ductile folds and fabrics in the country rocks, as first noted by Read (1931).

(b) Contacts between syenite and Moine or Lewisian country rocks are always sharp, with much veining of the country rocks adjacent to mapped contacts. Xenoliths are locally common, the largest (Moine psammities, measuring 500 x 300m approx.) occurring in the Ben Loyal intrusion at the southwest end of Bealach Clais nan Ceap (Map 4).

(c) The country rocks are nearly always strongly hornfelsed in the immediate contact zone (approx. 10m), but the exact limit of the hornfelsed zone as seen in the field, is difficult to define, since most of the country rocks are Moine psammities. Metasomatism is minor, and mainly restricted to the country rocks around the Cnoc nan Cuilean body, where metasomatic pyroxene and hornblende is commonly seen within a few metres of the contact (NC 5985 4547).

(d) Some of the garnetiferous semipelites lithologies around Ben Loyal locally develop abundant fibrolite after biotite (see Plate 6.2).

6.4.2 *Intrusion geometries observed in the field*

Previous workers have proposed a number of intrusion geometries for the Loch Loyal igneous complex. They were originally mapped as a single body by Horne during the first geological survey of the region (summarised by Geikie 1888). Phemister (1936, 1948) and King (1942) recognised that the three bodies were separate entities, but proposed differing models for the intrusion shape. Phemister suggested that the Ben Loyal intrusion has a sheet-like or laccolith form, whilst Ben Stumanadh was described as a series of irregular sheets running E-W, sub-parallel to the banding in the adjacent country rocks; Cnoc nan Cuilean was described as a 'stock'. King (1942) suggested an alternative, proposing that both Cnoc nan Cuilean and Ben Loyal had a conical form, with the apices pointing downwards and to the southeast. The most recent studies by Robertson & Parsons (1974) broadly concurred with the models suggested by Phemister, suggesting that the Ben Loyal intrusion has a wedge shape with a shallow-dipping roof to the southeast and steep, outward-dipping contacts to the north and west. They suggested that forceful emplacement of the syenite distorted and folded the regional foliation on the northern and western sides of the pluton.

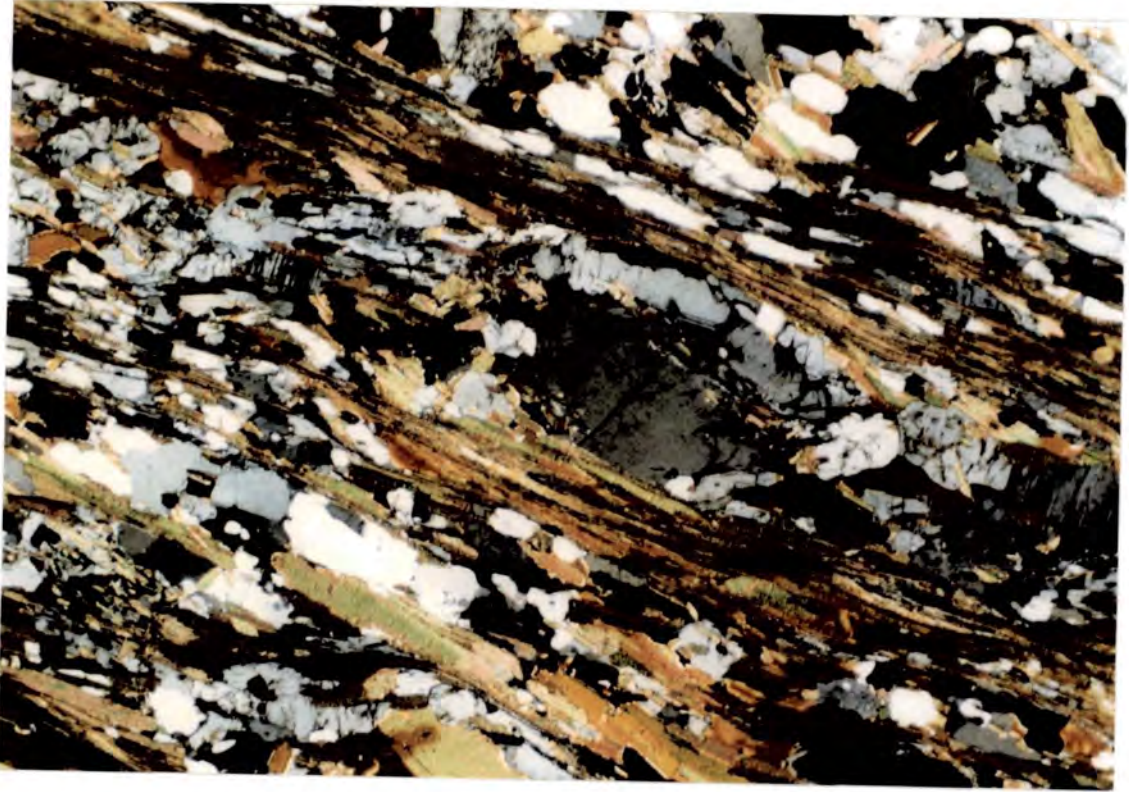


Plate 6.2 Photomicrograph of garnetiferous semipelitic country rocks (NC 5909 4922) surrounding the Ben Loyal Syenite. The photograph illustrates the development of fibrolite (dark needles) after biotite defining schistosity. (Field of view 4.4mm).

Figure 6.4 and Map 4 illustrate the geometry of the Loch Loyal complex in plan view based on remapping during the Sheet 114 contract and by the author. The Ben Stumanadh and Cnoc nan Cuilean bodies comprise a series of en echelon, variably coalesced NW-SE trending dykes, which usually run subparallel with the regional country rock foliation. Dykes occur on all scales from centimetres to tens of metres across.

The kinematics of intrusion can be determined in the field by examination of offsets of markers, drag folds and oblique fabrics in narrow dykes. Oblique fabrics and mineral lineations consistently indicate a dextral and SW-side down sense of motion, as seen by the NW plunging mineral lineation which is patchily developed throughout the three bodies. These features are summarised in Figure 6.7, and they all indicate that there is a small but persistent, component of dyke parallel dextral shear, with some additional extensional displacements occurring in places. Sub-vertical, NW-SE striking magmatic state deformation fabrics occur throughout both the Ben Stumanadh and Cnoc nan Cuilean bodies, lying at low-angles, or sub-parallel to the dyke margins (Map 4).

The Ben Loyal body has a more regular, semi-circular form. As mentioned in section 6.3, an internal, relatively unfoliated unit which grades into an outer, well-foliated facies. This becomes quite vuggy along very well-developed foliation planes, in a marginal facies adjacent to the contact with the country rock (see Map 4). Steep NW-SE trending magmatic state deformation fabrics, with or without aligned country rock xenoliths and mafic syenite, are common throughout most of the Ben Loyal body. In the NW of the body, however, the fabrics swing into approximate concordance with the contact (see Map 4). In the only fully exposed section across the contact, seen in Allt Chaonasaide, it appears that the dip of the foliation within the syenite decreases as one moves towards the contact and the psammitic country rocks here dip underneath the intrusion (see Fig. 6.8). The margin of the syenite intrusion is consistently steeper in dip than the country rock foliation on the north and northwestern sides of the pluton, but the dip is to the south and southeast, not north and west as reported by Robertson & Parsons (1974). In the extreme west, around Sgor Fhionnaich, the contacts are N-S trending and vertical, as reported by Robertson & Parsons (1974), but new exposures in the Allt Fhionnaich section show that the contact described here by these authors is, in fact, a WNW-ESE-trending fault (Map 4). In Allt Chaonasaide,

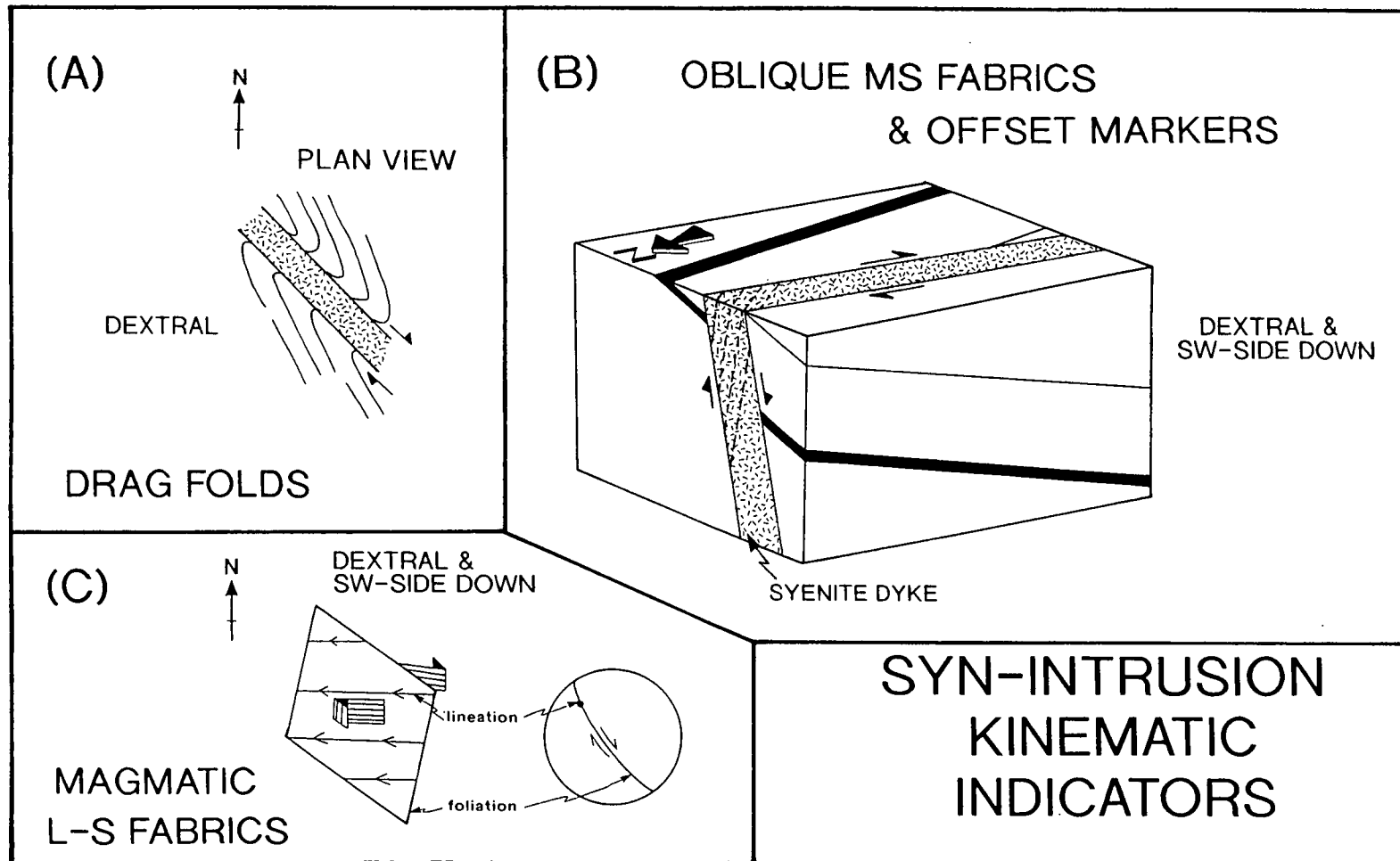
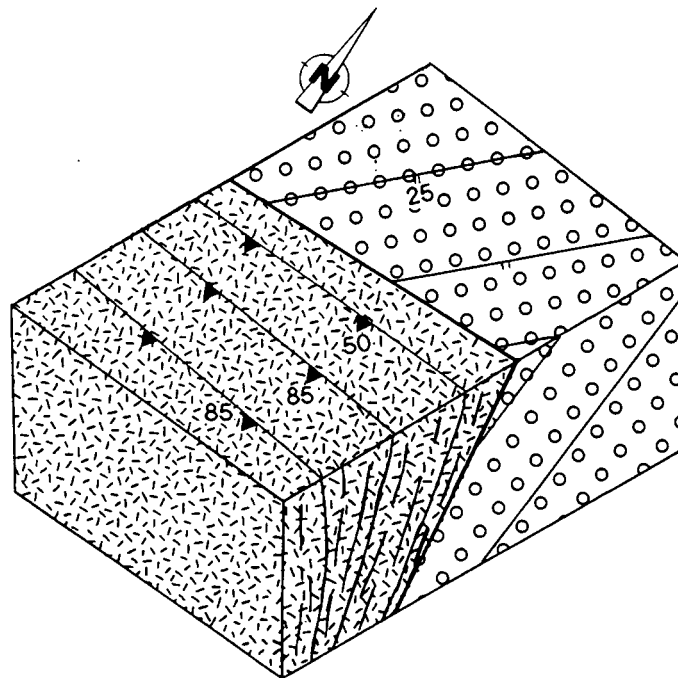


Figure 6.7 Syn-intrusion Kinematic indicators in the Loch Loyal Syenite Complex.

shallowing of the magmatic foliations in the syenite towards the contact gives an apparent overthrust sense of shear (Fig. 6.9). The present author interprets this apparent thrusting as a localised feature due to spreading of the magma to fill a cavity created by the interaction of regional tectonics, pre-existing structural architecture and magma intrusion (see section 6.5). This can be considered as a manifestation of 'weak' ballooning, in a manner akin to the intrusion mechanism postulated by Robertson & Parsons (1974).

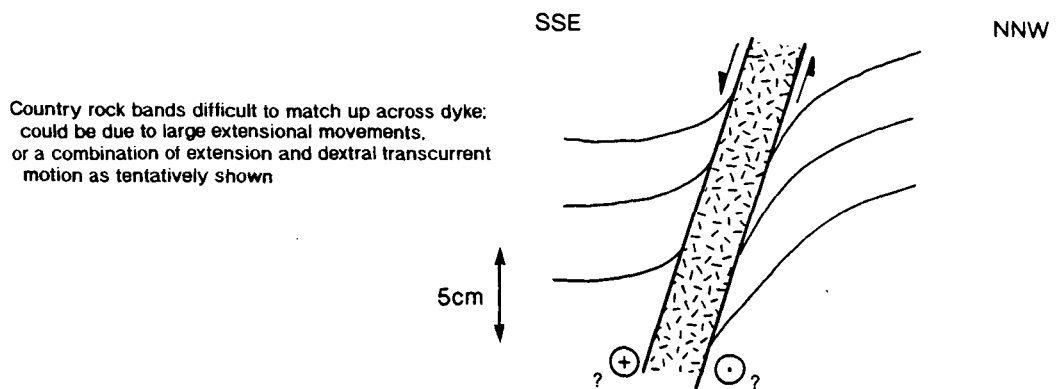
Figure 6.8 Sketch to show the relationship between the country rock foliation, the syenite contact and the orientation of the foliation in the syenite at the NW end of the Ben Loyal body in Allt Chaonasaide. Note that the contact dips more steeply and strikes clockwise to the country rock foliation. Also note that the syenite fabric shallows into the contact.



The kinematics of intrusion, as seen in narrow dykes of syenite intruded into the country rocks close to the N and W contacts (e.g. NC 57495012), does not indicate any evidence of regional overthrusting in association with syenite intrusion. Here, narrow dykes of syenite, like those around the Ben Stumanadh and Cnoc nan Cuilean bodies, show evidence of small to moderate amounts of extension; together with possible small wrench displacements (see Fig. 6.10 and section 6.5 for further discussion of kinematics of intrusion).

The south and eastern contacts of the Ben Loyal intrusion are very poorly exposed. In the region around Creag na Speireig, the syenite is clearly seen being emplaced as a series of northeasterly-inclined dykes lying sub-parallel to the regional foliation in the Moine and Lewisian country rocks (Map 4). Thus, the eastern (and higher?) part of the pluton appears to comprise a series of coalesced dykes with a form analogous to the Cnoc nan Cuilean and Ben Stumanadh (see also section 6.5).

Figure 6.9 Kinematics of intrusion of syenite sheets around the NW side of the Ben Loyal body, near Allt Chaonasaide.



6.4.3 Deformation fabrics associated with the emplacement of the Loch Loyal Complex

Microstructure was examined in a set of approximately forty thin sections, which were collected randomly throughout the main syenite bodies, and are representative of variations in both composition and deformation fabrics.

Most of the syenite units within the Loch Loyal complex, with the exception of the inner unit of the Ben Loyal syenite, carry well-developed deformation fabrics. Deformation apparently began when the syenite was still in its magmatic state (*sensu* this thesis), producing a strong shape preferred orientation (SPO) of mafic minerals and feldspar laths (see plates 6.3, 6.4, 6.5 & 6.7). The magmatic state fabric is variably overprinted by a solid state fabric throughout the pluton. Where it is developed, the solid state fabric is generally coaxial with the magmatic state fabric. An additional cataclastic fabric is locally

present, forming the lowest temperature fabric within the syenite complex. It is best developed in the Ben Stumanadh body adjacent to the shores of Loch Loyal, and can be related to late movements on brittle faults, in particular the Loch Loyal fault (Map 4). Since the individual bodies of the Loch Loyal complex exhibit slightly different deformation fabrics, the deformation history of each unit will be discussed separately.

Ben Loyal

The deformation history of the Ben Loyal body appears to be very different in the component units. The inner, or core syenite carries only a very weakly developed magmatic state deformation fabric, if any at all, whereas the outer unit is very well foliated; the intensity of the foliation increasing towards the contact with the country rocks.

The magmatic state fabrics can be observed in hand specimen and thin section. They are basically shape preferred orientations of pyroxenes, amphiboles where present, and feldspar laths (Plates 6.3 & 6.4), which trend NW-SE in the southeastern end of the Ben Loyal body, swinging into concordance with the contact of the pluton towards the NW end of the body.

The magmatic fabric is variably overprinted throughout the Ben Loyal body by a solid state deformation fabric which appears to be broadly coaxial with the magmatic state fabric, since it is parallel to it. The solid state fabrics are particularly well-developed around the NW contact zone. It is significant to note that many of the grain boundaries, particularly of quartz crystals, have serrated grain boundaries. This feature is typical of a rock that has undergone grain boundary migration, and secondary grain growth processes as significant deformation mechanisms. Such mechanisms require quite special conditions, for example they are common in rocks where the temperature is relatively high, because elevated temperatures facilitate the accommodation of higher strain rates. Furthermore, high water vapour pressures, for which there is circumstantial evidence in parts of the Ben Loyal body, favour the operation of grain boundary migration, secondary grain growth and diffusive mass transfer processes. It is interesting to note that the development of solid state fabrics is most pervasive nearest the northern and western margins of the unit, where the syenite has developed a slightly vuggy texture, believed to be due



Plate 6.3 The appearance of the magmatic state deformation fabric in the Ben Loyal Syenite. The fabric is defined by a strong shape preferred orientation of mafic minerals and feldspar phenocrysts.

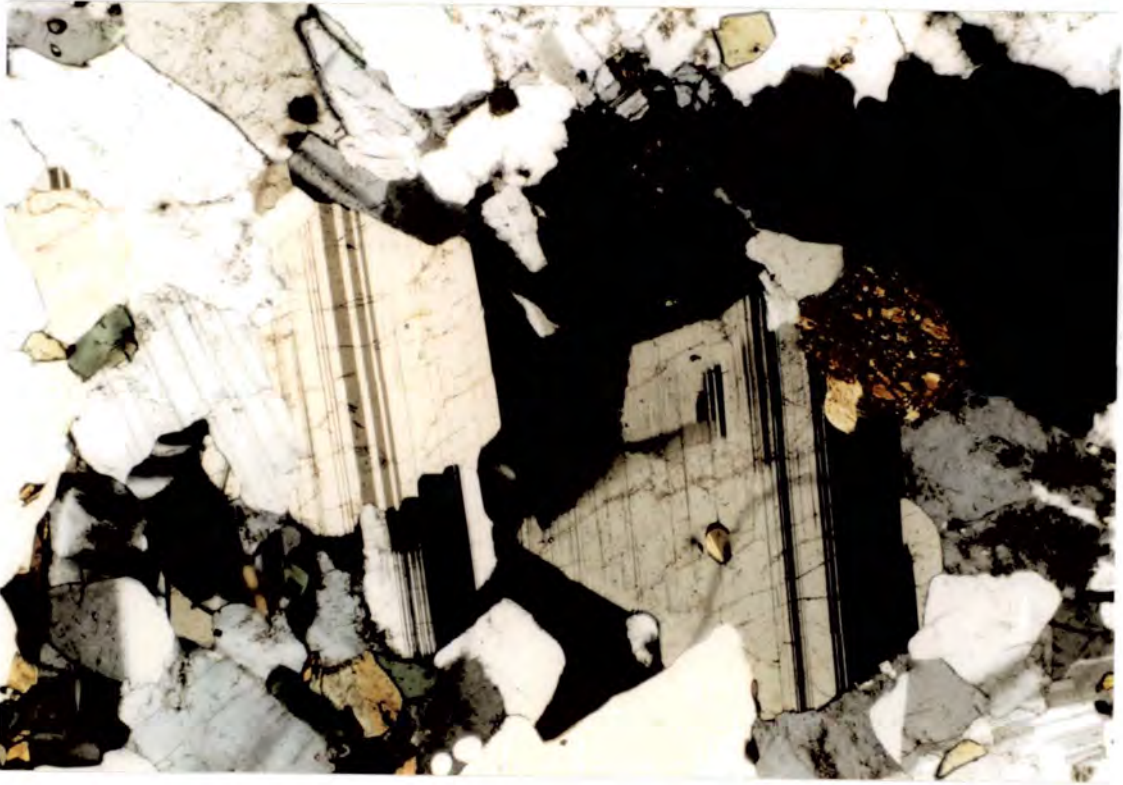


Plate 6.4 Photomicrograph of Ben Loyal Syenite. The photograph illustrates the presence of a strong shape preferred orientation of feldspar phenocrysts and mafic minerals set in a groundmass of equant quartz. (Field of view 4.4mm).

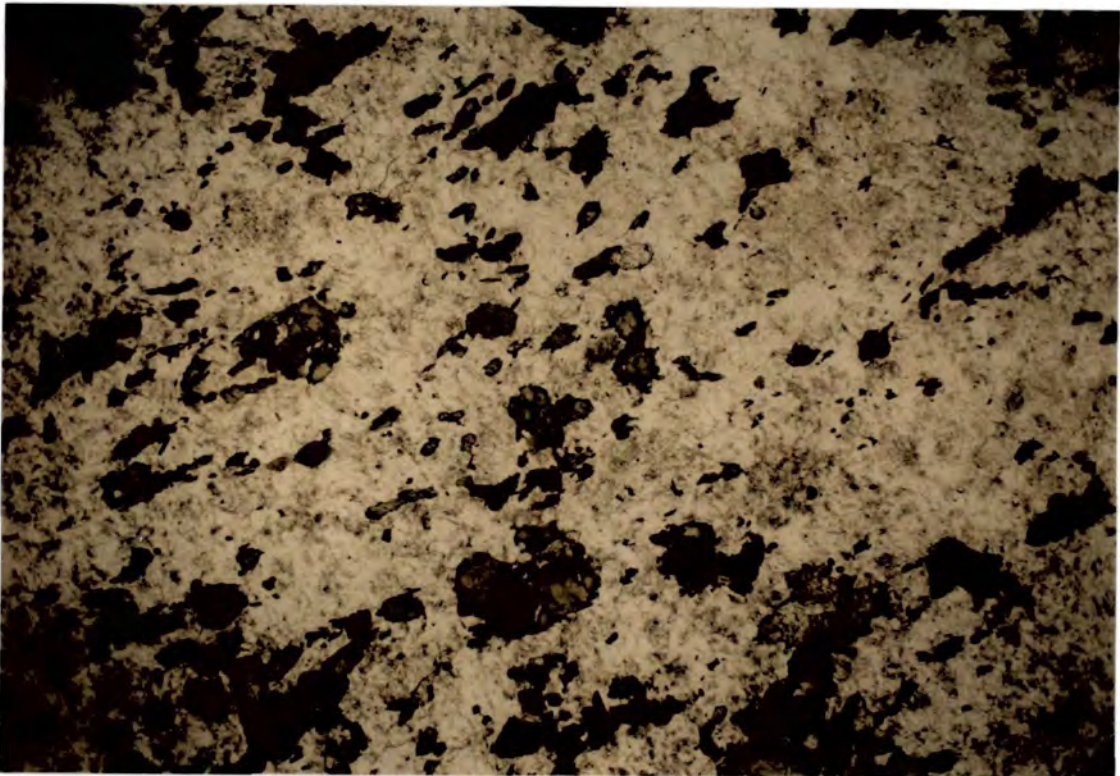


Plate 6.5 Photomicrograph of Cnoc nan Cuilean Syenite. Photograph illustrates the development of a strong shape preferred orientation of feldspar phenocrysts defining a well developed magmatic state deformation fabric. (PPL; field of view 18mm).

to high water vapour pressure (Robertson & Parsons 1974). This together with the 'semi-annealed' nature of the groundmass in many of the Ben Loyal rocks suggests that the solid state deformation must have occurred either when the magma was still relatively hot, or in association with high water vapour pressures, or a combination of both of these physical parameters. If a water gradient existed within the pluton, possibly due to fluxing of water between the syenite and the country rocks, but resulting in the highest water vapour pressures being found near the margins of the pluton. The water gradient resulted in the production of a two feldspar assemblage, as seen in the outer facies, from a one feldspar assemblage, as seen in the inner facies (Robertson & Parsons 1974).

Cnoc nan Cuilean

Like Ben Loyal, Cnoc nan Cuilean carries a well-developed shape preferred orientation of pyroxene and feldspar laths, typical of a magmatic state deformation fabric (Plate 6.5). The intensity of the deformation fabric varies throughout the body, since the Cnoc nan Cuilean body is highly variable in composition, a feature which is believed to reflect the fact that the intrusion is composed of a series of coalesced dykes of variable mineralogical composition. On initial inspection, the magmatic state fabric appears to be strongest in the more basic bodies. However, careful analysis in the field and in thin section reveals that this is not real, but a form of optical illusion, because it is much easier to recognise the fabric in the more basic units, since the SPO is mainly defined by mafic minerals, whereas, in more felsic sheets the fabric is defined by mafic minerals and aligned feldspars, that do not 'attract the eye' as readily. The nature of the magmatic state deformation fabric is well illustrated in thin section (Plate 6.5). However, the magmatic state deformation fabric in this body, like Ben Loyal, is often overprinted by a relatively high temperature solid state deformation fabric, which produces subgrain development around feldspar phenocrysts, some recrystallisation of quartz and feldspar and a semi-annealed texture (Plate 6.6). The relatively limited development of the solid state fabrics and the semi-annealed texture suggests that solid state deformation in this unit also occurred at relatively high temperatures and did not involve large amounts of strain.



Plate 6.6 Photomicrograph of a specimen of Cnoc nan Cuilean Syenite illustrating the development of a high temperature solid state deformation fabric overprinting a magmatic state deformation fabric. The solid state deformation results in subgraining and primary recrystallisation around the margins of feldspar phenocrysts and recovery and recrystallisation of groundmass quartz. Note also the relative abundance of mafic minerals in this syenite. (Field of view 18mm).



Plate 6.7 Photomicrograph of Ben Stumanadh Syenite illustrating the occurrence of a well developed magmatic state deformation fabric. The fabric is defined by a shape preferred orientation of feldspar laths and mafic minerals. There has been some degree of solid state deformation in this specimen also. (Field of view 18mm).

Ben Stumanadh

Ben Stumanadh, like the two other bodies, has a well-developed magmatic state deformation fabric, defined by aligned feldspar laths and prismatic mafic minerals (Plate 6.7). In contrast to the other two bodies, there is little evidence of high temperature solid state deformation fabrics, or semi-annealed textures in the Ben Stumanadh syenite. In most samples, solid state deformation features are characteristic of deformation at lower temperatures ($\leq 450^{\circ}\text{C}$ from brittle-ductile behaviour of feldspars; Simpson 1985), since they have suffered crystal plastic strain. The microstructures formed in feldspar appear rather brittle in nature, producing kinking and sometimes slight fracturing of feldspar (Plate 6.8). Recovery and recrystallisation features, such as subgrain development and primary recrystallisation are common in quartz (Plate 6.9). In addition, graphic intergrowth of quartz and feldspar is sometimes observed within the syenites of this body, a feature which is often indicative of crystallisation at higher crustal levels, because the physical conditions near the surface result in quartz and feldspar reaching their solidus simultaneously. Vuggy textures are very common, and although some may be related to late brittle faults (see below), the Ben Stumanadh body is much vuggier than most of the Ben Loyal body, with the possible exception of the marginal phase of the latter. This, along with the petrographic evidence outlined in section 6.3, suggests that the Ben Stumanadh body may be a higher level of exposure of the Ben Loyal body. Its textural characteristics and deformation history are slightly different to the Ben Loyal body because of relatively high water vapour pressures, and intrusion closer to the surface than Ben Loyal.

6.4.4 Late deformation of the Loch Loyal Complex

The last phase of deformation observed in the syenite complex is associated with movement along late brittle faults, in particular the NE-SW striking major fault which runs through Loch Loyal (see Fig. 6.4 & Map 4), a structure which will be named the Loch Loyal Fault in the new memoir on the area (Holdsworth *et al.* 1993). This fault is believed to have formed in relation to the offshore development of the West Orkney Basin (Holdsworth 1987, 1989; Enfield & Coward 1987). Kinematic indicators along the fault indicate that the sense of movement was dextral, with a component of downthrow of the southeastern block. Brittle deformation associated with fault movement mainly affects the Ben



Plate 6.8 Photomicrograph of Ben Stumanadh Syenite showing dextral offset in feldspar phenocryst due to brittle fracturing. (Field of view 4.4mm).

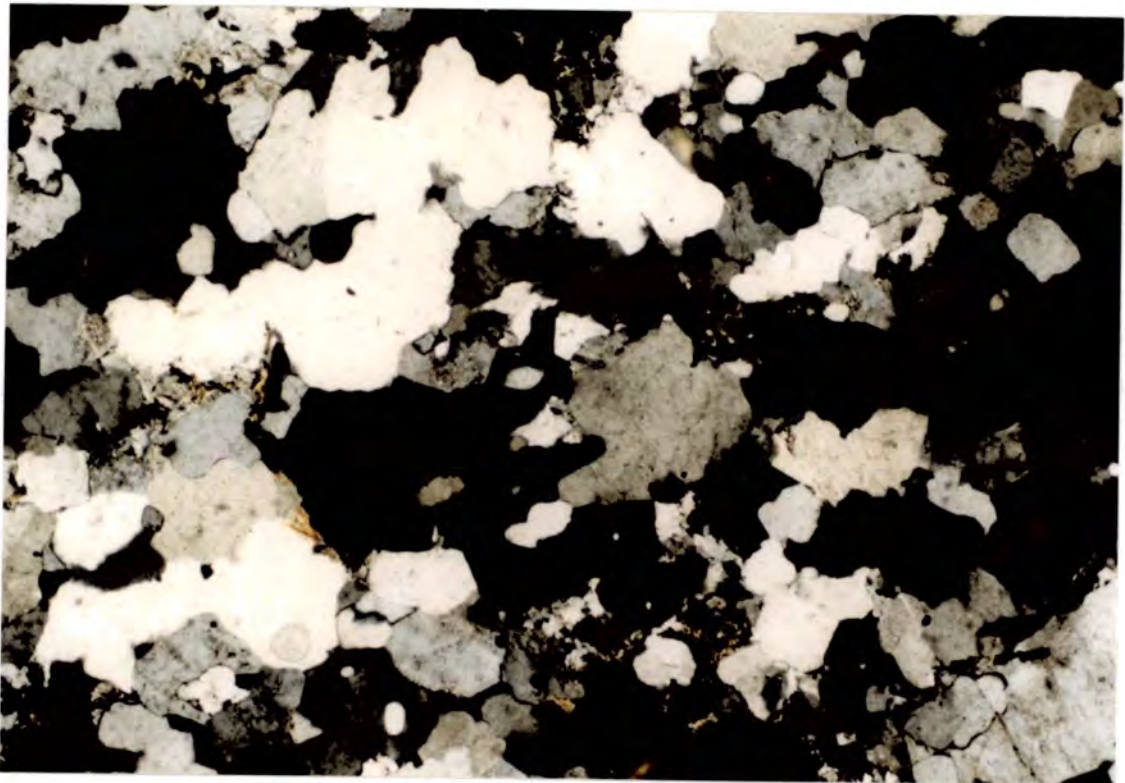


Plate 6.9 Photomicrograph of Ben Stumanadh Syenite illustrating the occurrence of recovery and recrystallisation features, such as recrystallisation and secondary grain growth, in groundmass quartz. (Field of view 4.4mm).

Stumanadh body, which, close to the fault is characteristically reddened (presumably due to Fe_2O_3), forming the vivid red screes and outcrops which give Sron Ruadh, above Loch Loyal its name (see Map 4). This red colour may also explain why some authors (e.g. O'Reilly 1971) have considered Ben Stumanadh to be more 'granitic' than the other bodies. The Loch Loyal Fault produces cataclastic breccias in a zone at least 50m across. The effects of cataclasis and brecciation can be seen in the field, but they are also particularly well-illustrated on a small scale in thin section (Plates 6.10). Many of the small scale fractures show a synthetic sense of movement to the main fault, that is dextral (Plate 6.10). A 5m zone within the fault breccia is mineralised (Plate 6.11). This contains texturally zoned carbonate that initially grew parallel to the walls of the zone, but which later grew as fine grained 'pockets' within the main mineralised zone. Hydrothermal fluids also percolated out through mesoscopic and microscopic fractures, precipitating along cracks and in vugs (Plate 6.12). It is believed that in addition to the primary vugginess in the Ben Stumanadh syenite, a secondary vugginess developed in response to fluid circulation in the fault zone, and was brought about by the breakdown of groundmass mafic minerals, such as biotite. It is these vugs that are commonly in-filled with carbonate material.

The nature of the mineralised zone suggests that it is an explosion breccia, implying that high fluid pressures existed along the fault zone in Devonian times. It is suggested that high fluid pressures were also associated with the earliest movements along this fault, resulting in the development of a zone of softening in the Ben Stumanadh syenite, producing relatively intense crystal plastic deformation of the syenite in a narrow zone close to the fault zone. This produces annealed recrystallisation textures, that are otherwise absent from the Ben Stumanadh body, indicated by the presence of features such as serrated grain boundaries, secondary recrystallisation and grain growth (see Plate 6.13). There is no indication that this fault was active during magma emplacement. It is thought unlikely that this deformation fabric is a down-temperature continuum of the emplacement-related fabrics, since it is not observed anywhere else in the Ben Stumanadh body, or the syenite complex as a whole.

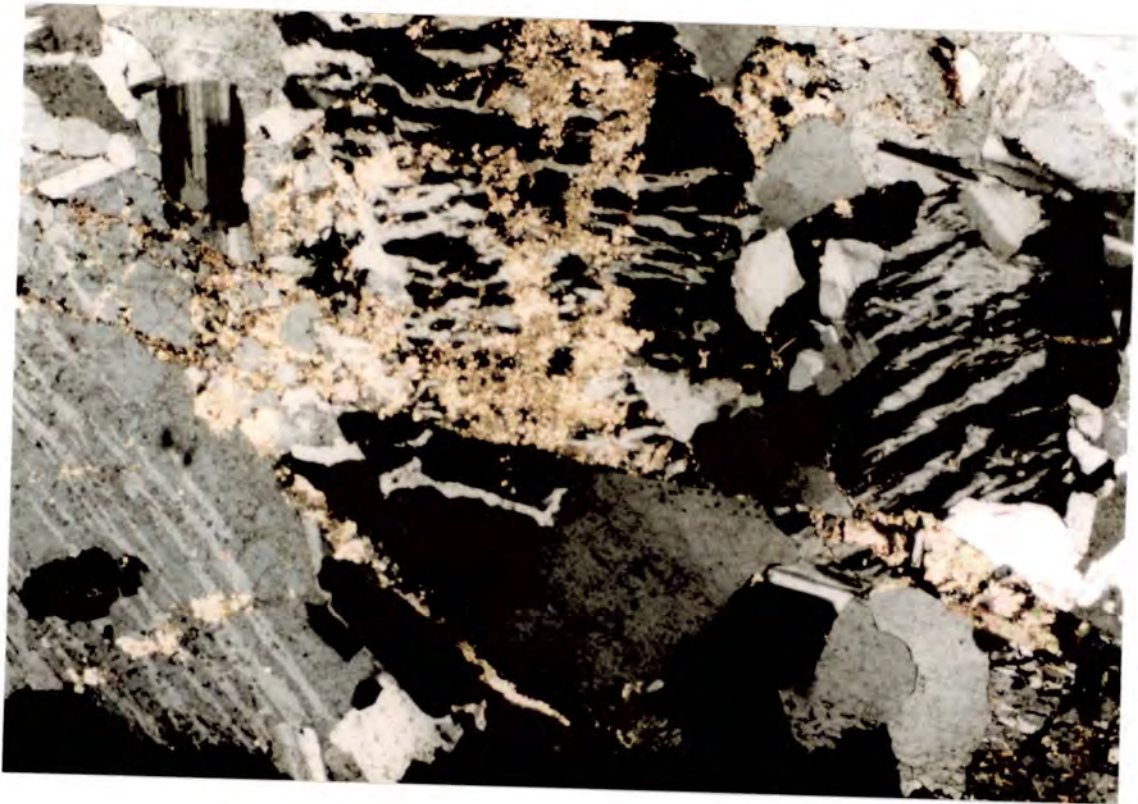


Plate 6.10 Photomicrograph of Ben Stumanadh Syenite illustrating the pervasive nature of cataclasis related to brittle movements along the Loch Loyal Fault in the syenitic wallrocks adjacent to the fault. Note that most cracks, fractures and vugs are now filled with fine-grained carbonate. (Field of view 4.4mm).



Plate 6.11 Appearance of the mineralised part of the Loch Loyal Fault Zone as exposed along the eastern shores of Loch Loyal.



Plate 6.12 Photomicrograph of Ben Stumanadh Syenite illustrating the presence of fine-grained carbonate material in vuggy cavities (bottom left). This material is believed to have precipitated from fluids percolating out from the fault zone. (Field of view 1.9mm).



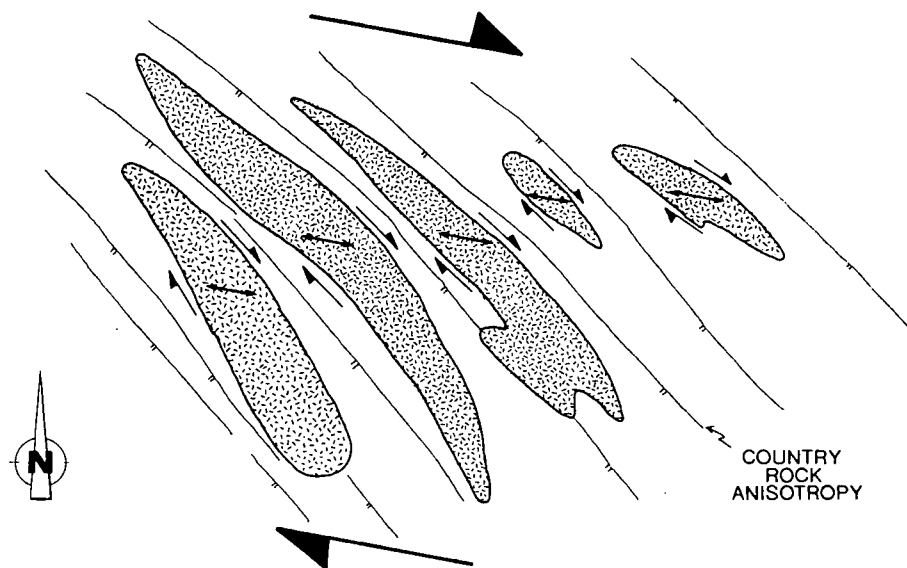
Plate 6.13 Photomicrograph of Ben Stumanadh Syenite from adjacent to the Loch Loyal Fault Zone. Note the polygonal texture that is believed to have been produced by the operation of crystal plasticity in response to elevated pore fluid pressures, which resulted in dynamic recovery and recrystallisation. (Field of view 18mm).

6.5 A model for the emplacement of the Loch Loyal Syenite Complex

6.5.1 Two dimensional intrusion geometry

The full emplacement model will be outlined later in this section, but it is proposed that the dykes forming the Ben Stumanadh, Cnoc nan Cuilean and eastern part of Ben Loyal intrusions formed as transtensive pull-aparts due to the interaction of a dextral shear couple with the unusually orientated, NW-SE trending country rock foliation.

Figure 6.10 Cartoon sketch summarising the development of transtensive pull-aparts due to the interaction of a dextral shear couple with NW-SE orientated country rock foliation to produce dextral pull-aparts with a small dyke-parallel component of dextral shear.



(PROCESS OBSERVED ON ALL SCALES)

Such an interaction would produce dextral pull-aparts and small dyke parallel dextral shear components. The scale of intrusion would be dictated by the magnitude of the strain and may be directly influenced by the spacing of the most incompetent units of country rock, as schematically portrayed in Figure 6.10.

6.5.2 *Three dimensional intrusion geometry*

The occurrence of the Loch Loyal fault is very fortuitous in terms of deducing the three dimensional geometry of the syenite complex, because it apparently exposes two different levels of the same intrusion, the Ben Loyal syenite at a lower structural level and the Ben Stumanadh syenite at a higher structural level. This is supported by the following observations.

(1) Displacement indicators along the exposed section of the fault (dextral and SE-side down).

(2) The overall petrography and composition of the units is not significantly different.

However, the Ben Stumanadh body exhibits some graphic intergrowth of quartz and feldspar, which could be indicative of emplacement and crystallisation at higher crustal levels. Furthermore, the occurrence of vuggy textures are ubiquitous throughout this body, whereas, in Ben Loyal they are confined to the most marginal zones (see subsection 6.4.2).

(3) The textural features of the bodies suggest that they have had slightly different deformation histories. The Ben Stumanadh body suffered a much lower temperature solid state deformation than the Ben Loyal body during, or shortly after emplacement.

(4) Both show evidence of dextral transtension.

All of the above observations are consistent with the Ben Stumanadh body being emplaced and deformed at shallower crustal levels than the Ben Loyal body. The three dimensional shape of the Ben Loyal-Ben Stumanadh syenite complex can, therefore be envisaged as approximating to the form portrayed schematically in Figure 6.11. The higher structural level of the pluton, the Ben Stumanadh body, has the form of an echelon transtensive dykes, and it is suggested that these 'pond' or coalesce downwards into a composite body, the Ben Loyal body, at depth. A similar coalescence is seen at the southwestern end of the Ben Loyal body. The form of the composite body at depth appears to be controlled by the pre-existing orientation of the country rock foliation, in the manner envisaged in Figure 6.12. The early structural architecture is mainly controlled by the development of the zone of F_3 cross folds, some of which run directly through the igneous intrusions. In particular, the Ben Stumanadh synform runs directly through the middle of the syenite on Ben Stumanadh (see Figs. 6.13 & Fig. 6.12). However, the pattern is slightly complicated around the NW of the Ben Loyal body, where syenite

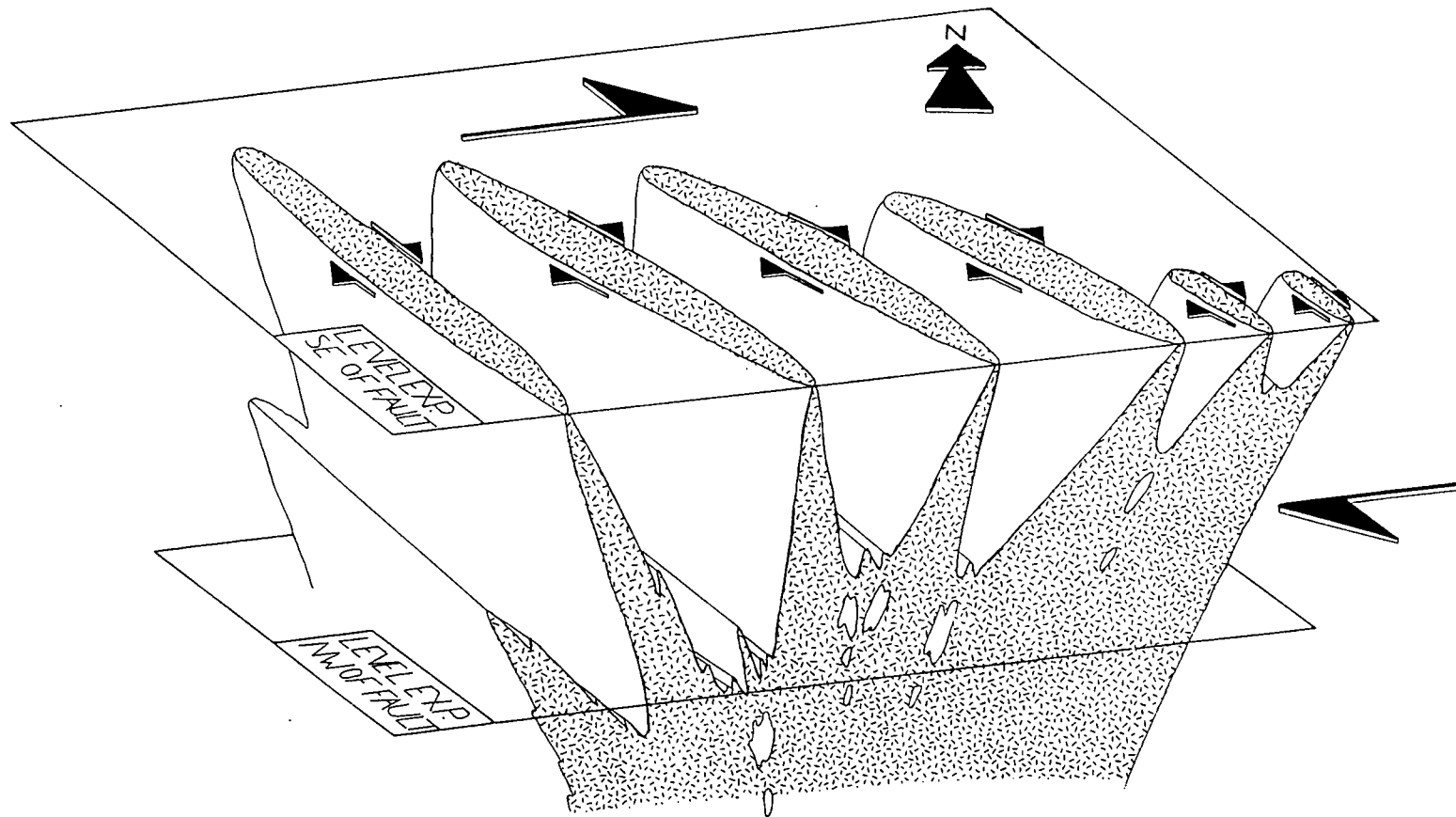


Figure 6.11 3-D model for the Ben Loyal - Ben Stumanadh composite body. Movement on the NE-SW fault exposed different levels of the same intrusive unit.

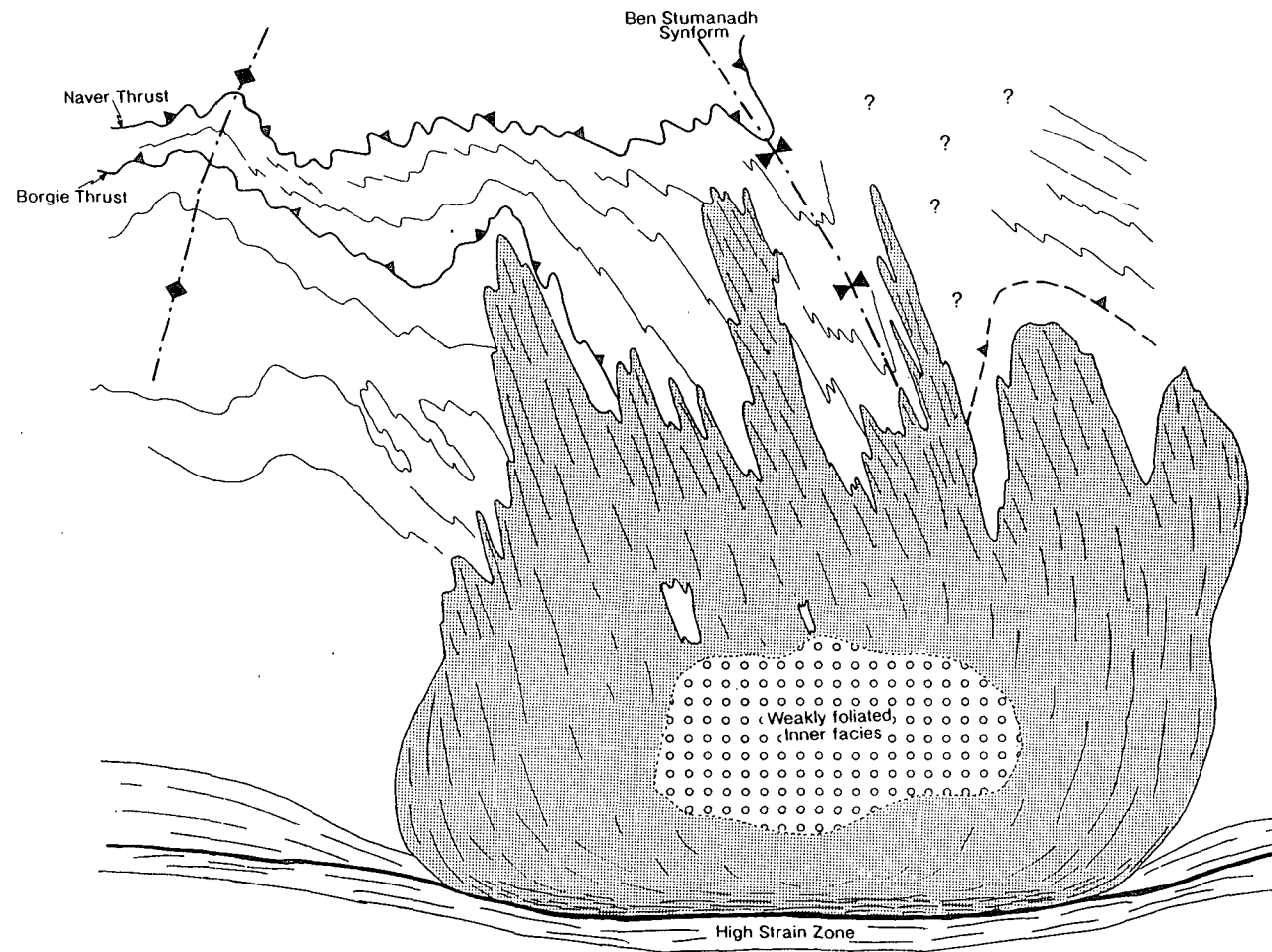


Figure 6.12 Schematic longitudinal cross-section parallel with the regional strike to illustrate the relative positions of ductile thrusts, crossfolds and the geometry of the Ben Loyal and Ben Stumanadh syenites prior to the formation of the Loch Loyal Fault.

emplacement appears to be influenced by a pre-existing regional high strain zone in the Moine country rocks, dipping gently towards the SE. This feature appears to be related to the cross folds, which appear to root into it (Fig. 6.12; R.E. Holdsworth pers. comm.).

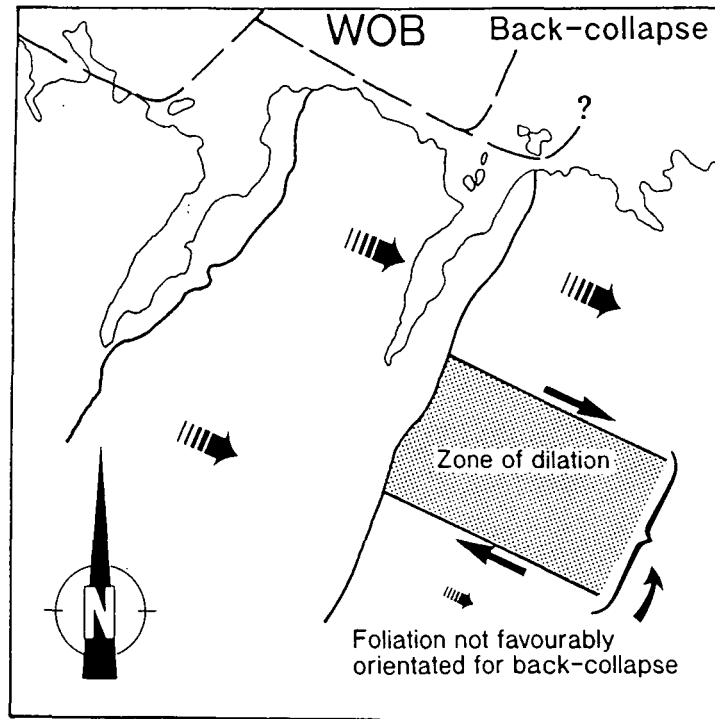
The three dimensional form of the Cnoc nan Cuilean body is less clear. All the petrographic, compositional and textural evidence points towards this being a separate, though related, intrusion from the other two bodies, but lack of exposure and the occurrence of the Loch Loyal fault in this area means that it is impossible to state this with total confidence. However, it also appears to be composed of a series of coalesced dykes, the margins of which can be located by variations in modal composition that can be crudely 'eyeballed' in the field. Support for a dyking model is observed around the NW of Cnoc nan Cuilean, SE of the late fault (Map 4 NC 5946) where sub-vertical syenite dykes, striking NW-SE, protrude into the country rock from the main Cnoc nan Cuilean body.

6.5.3 Regional tectonics and magmatism

Holdsworth (1989) has discussed evidence for ESE-directed back-collapse of the Moine Nappe under the influence of gravitational forces, late in the Caledonian orogenic cycle. This is facilitated by the NNE-SSW orientated ductile thrust fabric developed over much of Northern Sutherland. However, the zones of cross folding, including that around the Loch Loyal syenite complex, are anomalous in that they are zones of steeply dipping, NW-SE orientated strata.

Thus, the disposition of strata in these zones is less favourably orientated for back-collapse and may have acted as a zone of sticking (see Fig. 6.14), resulting in the development of a dextral shear couple across the zone of cross-folds which, leading to failure along regional foliation planes. Magma was emplaced as a series of en echelon, dextral transtensive dykes, that carry well-developed NW-SE trending deformation fabrics. These coalesce at depth into a more coherent body, the Ben Loyal unit. The deformation fabrics in the south of this body are sub-vertical and NW-SE trending, which is similar to those described from the discrete en echelon dykes, so a similar kinematic setting seems likely.

Figure 6.14 Schematic diagram showing how the zone of NW-SE banding may have formed a dextrally transtensive dilation zone, favouring igneous emplacement during Caledonian gravitational back-collapse (c. 426Ma). Note that the margins of the zone of dilation are parallel to the presumed orientations of transfer faults forming the southern boundary of the offshore West Orkney Basin. This trend may reflect a NW-SE structural grain in the (Lewisian) deep basement.



However, around the north and west margins of the Ben Loyal body the deformation fabrics swing in to concordance with the moderately dipping contacts above the early high strain zone. This fabric pattern in the syenite is believed to reflect 'ballooning', or spreading of the magma to fill a saucer-like cavity, created due to the coincidence of the cross folds and a SE dipping high strain zone around the NW margin of the intrusion (see Fig. 6.12).

These observations suggest to the present author that gravitational collapse permitted magma to be emplaced into the Moine Nappe at approximately 426Ma. Magma emplacement occurred in areas where the structural grain of the country rocks was unfavourable for back-collapse to occur along the dip of the foliation, so that a dextral transtensive stress field developed, resulting in the opening of en echelon, transtensive pull-aparts into which magma could be emplaced.

Conclusions

1. Intrusion of the Loch Loyal syenite complex *c.*426Ma (U-Pb; Halliday *et al.* 1986) began with the emplacement of the Cnoc nan Cuilean body. The form of this pluton resembles a group of NW-SE orientated, coalesced dykes of variable modal composition. It is the most basic and alkalic of the Loch Loyal syenites.
2. This was followed very closely by emplacement of the Ben Loyal and Ben Stumanadh Syenites. These bodies appear to represent different structural levels of the same pluton, since their petrography and chemical composition are virtually identical. Ben Stumanadh represents shallower structural levels than Ben Loyal. The former is a series of en echelon, dextrally transtensive dykes, which passed down into the more coherent Ben Loyal body.
3. Variation in intrusion level is reflected by differences in the textures of the units, as a result of slightly different deformation histories in response to different physical conditions, such as pressure, temperature and water vapour pressure.
4. Different levels of the same pluton are exposed due to the chance occurrence of the Loch Loyal Fault, which cuts the syenite complex. The motion on this fault, believed to be related to the formation offshore of the West Orkney Basin, is dextral, with a significant component of downthrow of the eastern block.
5. Emplacement of the syenite complex post-dates major ductile thrust displacements within the Moine Nappe. However, syn-thrusting differential displacements generated large-scale cross-folds that have subsequently strongly controlled the siting and mode of intrusion of the syenites.
6. Gravitational collapse of the nappe pile caused dextral transtension in the region of the earlier cross-folds, thereby creating space for magma emplacement to occur.

Chapter 7

DISCUSSION AND GENERAL CONCLUSIONS

7.1 Pluton shape and mode of emplacement

The three plutons described in this thesis were all emplaced under conditions of transtension, created by the interaction of the pre-existing structural architecture and strike-slip-dominated regional tectonics at the time of emplacement. The importance of the pre-existing country rock structure in controlling the siting and geometry of intrusive bodies has been overlooked by recent workers, with most studies tending to concentrate on the effect of tectonic processes (e.g. Hutton 1988 etc.). There can be little doubt that tectonic controls may dominate in some circumstances, especially in areas where the strains associated with pluton emplacement are large, i.e. shear-zone hosted intrusions (Main Donegal Granite, Great Tonalite Sill, St Malo Migmatites). However, the Caledonian plutons studied in the course of this work *do not* appear to have been associated with particularly large strains. In such cases, the evidence presented in this thesis demonstrates that regional tectonics alone cannot account for the observed shapes, fabric evolution and modes of pluton emplacement; this point has recently been alluded to by Paterson and Fowler (1993). Instead, these features are more directly controlled by the pre-existing structural architecture in the country rocks. Thus, whilst tectonic forces act as the catalyst in initiating the creation of space into which magma can be emplaced, the pre-existing structural architecture will control the ultimate form of the pluton.

7.2 The evolution of deformation fabrics during pluton emplacement

Deformation fabrics within the three plutons are dominated by magmatic state fabrics. This may also be a reflection of relatively small and localised tectonic strains in association with magma emplacement. However, in the case of the Ratagain Complex, there is some evidence to suggest that deformation is partitioned into the least competent ('softest') units. This is significant when examining plutons where there is evidence for more than one pulse of magma (e.g. sheeted plutons such as the Main Donegal Granite and the Great Tonalite Sill), since the absence of solid state deformation fabrics from certain intrusive bodies could be interpreted as indicating that there was a cessation in regional deformation. Whilst it may be true to say that deformation was not continuous

throughout the crystallisation history of individual intrusive units, it may not be valid to postulate that regional deformation had completely ceased, since it may simply have been partitioned into a later pulse of magma.

In the cases of the Thorr and Loch Loyal plutons, there is evidence that the production of coplanar solid state deformation fabrics is a localised phenomenon that may be related to the presence of hydrous fluids. Such conditions could lead to hydrolytic weakening of minerals such as quartz and biotite and ultimately to a focusing of strain into these weakened zones.

Magmatic state deformation fabrics in the three plutons are often aligned parallel to the intrusion margins. Since there was no evidence for high shear strains, the author accounted for such patterns in the Thorr Pluton in terms of the production of a buoyancy head that develops in response to the body forces that are exerted by the magma acting across the intrusion walls (Fig. 7.1).

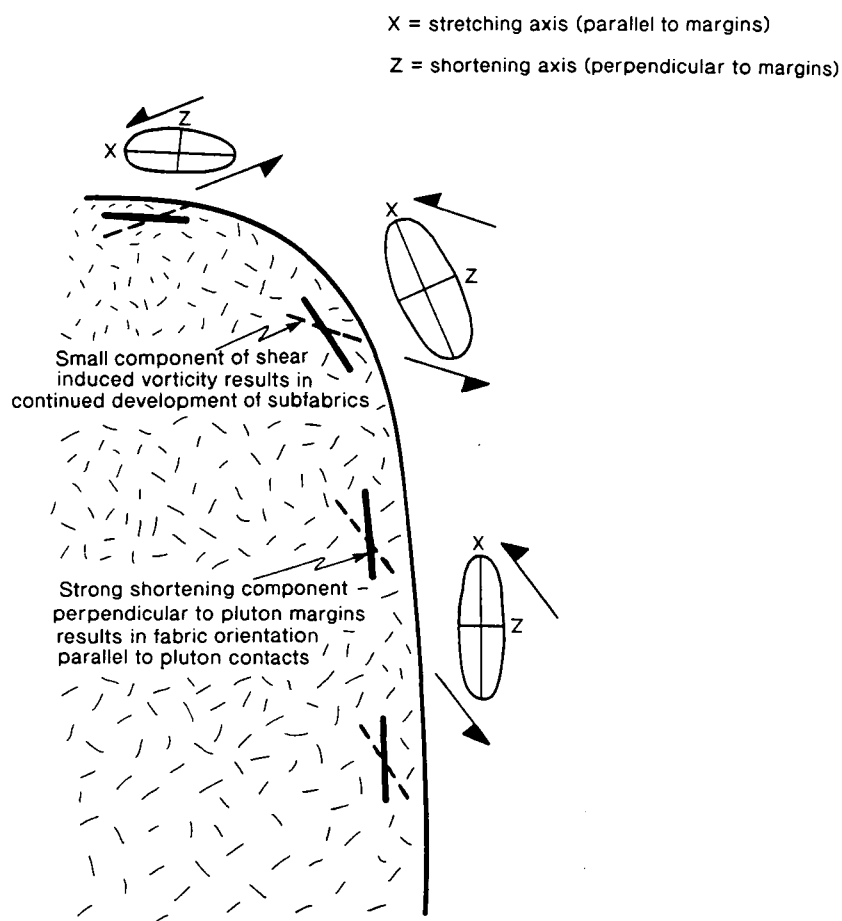


Figure 7.1 Cartoon sketch to illustrate the development of 'buoyancy head' fabrics. These are produced as a result of an internal component of pure shear in response to the magma acting across the intrusion walls. Shear arrows indicate the component of shear induced vorticity within the intruding magma.

Similar fabrics are developed in the Ratagain and Loch Loyal plutons, but in these plutons magmatic state fabrics generally have only one preferred orientation of phenocrysts that is often aligned parallel to the intrusion walls.

A simple explanation of such fabric geometries is that the body forces exerted by the magma across the intrusion walls produce a localised coaxial strain component in the magmatic state, in which the axis of shortening is normal to the pluton walls. The author believes that fabrics produced in this way are *not* the same as magmatic flow fabrics (Balk 1937), since this implies that such fabrics are internally created in response to flow of the magma past the rigid walls. However, flow is not required in the formation of buoyancy head fabrics. The implication is, therefore, that when examining deformation fabrics, one must study their geometry and the state of strain in detail before attributing their formation solely to the operation of external tectonic stresses during, or immediately after emplacement.

The method of Fernandez and Laporte has been shown to work in the field. The results obtained from application of the technique in the Thorrr pluton consistently indicate sinistral sense of shear from the magmatic state fabrics within the granite (Fig. 7.1); this is consistent with kinematic analyses along the margins of the pluton. However, the fact that some of the magmatic state fabric have clearly developed under conditions of very low strain indicates that the development of foliations and fabrics, especially in the magmatic state, does not require large amounts of tectonic strain and/or can develop in response to internally generated body forces. The technique of Fernandez and Laporte does not distinguish the origin or magnitude of the strain that results in the development of magmatic state subfabrics; such information can only be derived from complimentary studies of contact relationships and strain analysis (see Ch. 4). This illustrates that, like all geometric and kinematic analytical techniques, the method of Fernandez and Laporte should not be applied in isolation.

7.3 Basement control on magma ascent: regional zones of differential shear

In all three study areas, a relationship has been established between the occurrence of strike-slip shear zones, or zones of differential displacement, and magma emplacement. It is possible that this reflects two fundamental features of strike-slip shear zones and faults. Firstly, their frequent

association with magmatism perhaps reflects their deep-seated nature, which allows them to tap into areas of magma production (D'Lemos *et al.* 1992; Hutton & Reavy 1992). Alternatively, the association between magmatism and strike-slip tectonics may reflect the ease of reactivation of such faults (Sibson 1983). Indeed, it has been suggested in this thesis that in Thorr and Ratagain, the emplacement of magma to its present level was triggered by the rheological weakening of strike-slip zones caused by magma appearing at depth; this would account for the otherwise unexplained renewed movement along these shear zones. It may be that like any other fluid, magma plays a fundamental role in 'lubricating' fault zones, resulting in reactivation. The interaction between strike-slip shear zones and magmas may follow a cyclical pattern, with magma triggering reactivation along deeply penetrating faults which can then provide active channelways for the syn-tectonic ascent and syn- to post-tectonic emplacement of these magmas at higher structural levels. Thus plutons such as Thorr, Ratagain and Loch Loyal may be underlain by sheeted complexes that are associated with higher tectonic strains.

The plutons studied in this investigation can, therefore, be considered to represent an intermediate stage between the sheeted intrusions that are emplaced in association with active shear zones, for example, the Main Donegal Granite, and passively emplaced, high level plutons such as the Rosses Granite.

7.4 Further Research

In the light of the conclusions of this thesis, the author makes the following recommendations for further work.

(1) It has been suggested here that not all deformation fabrics are the product of externally applied stresses, but that some magmatic state fabrics may result from internal body forces within plutons. If this is generally the case, then it is important to develop a method by which 'buoyancy head' fabrics can be distinguished from magmatic state fabrics produced in response to externally applied stresses. An obvious way of doing this is to relate the geometry and evolution of the deformation fabrics to strain and external kinematic data, where available. It may also be possible to model the development

of such fabrics in the laboratory using magma analogues, such as those used by Fernandez and Laporte (1991) in a rigid container.

(2) It is useful to examine in detail the distribution of granite fabrics (i.e. magmatic state and solid state) within plutons in order to determine whether there is significant partitioning of strain according to such factors as (a) order of intrusion, (b) composition of intrusive unit (i.e. in terms of chemistry, mineralogy and crystal-melt ratios), (c) the presence of primary strain variations due to the location of major shear zones, (d) composition and degree of mobilisation of wall rocks. Furthermore, does the absence of solid state fabrics indicate a cessation in regional deformation, or can it be related to spatial and temporal partitioning of strain? All of these factors are important in determining the relative structural age of granite fabrics and have direct bearing on the interpretation of isotopic dates from plutons, which are commonly used to constrain regional geological histories.

(4) The technique of Fernandez and Laporte should be applied in a wide range of tectonic settings in order to truly test its applicability and validity. It is also important to compare the results from the different tectonic settings to determine whether there are any fundamental differences in the evolution of magmatic state deformation subfabrics in shear zone-hosted, intermediate and relatively passively emplaced intrusions.

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APPENDICES

APPENDIX A

LOC.1525	Subfabric A Low A.R.				Subfabric B High A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO	
20	15	47	1.33	19	6	110	3.17	
21	15	42	1.40	30	10	77	3.00	
25	14	48	1.79	26	7	87	3.71	
19	12	16	1.58	24	6	80	4.00	
22	13	54	1.69	23	7	83	3.29	
16	11	49	1.45	17	5	82	3.40	
19	13	33	1.46	15	5	55	3.00	
21	12	38	1.75	16	6	56	2.67	
33	18	51	1.83	20	6	88	3.33	
19	14	40	1.36	26	6	115	4.33	
25	17	37	1.47	27	8	74	3.38	
16	11	40	1.45	12	5	67	2.40	
20	15	10	1.33	22	7	67	3.14	
24	14	40	1.71	26	7	75	3.71	
18	11	45	1.64	30	5	65	6.00	
17	6	60	2.83	12	5	52	2.40	
20	15	40	1.33	18	6	86	3.00	
17	11	46	1.55	12	5	95	2.40	
26	17	36	1.53	25	6	114	4.17	
15	10	30	1.50	22	4	118	5.50	
AVERAGE = 040/86NW				AVERAGE = 086/84N				

APPENDIX A

LOC.1548				LOC.1548			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
15	13	124	1.15	12	3	176	4.00
11	8	144	1.38	28	8	133	3.50
15	12	168	1.25	30	9	6	3.33
13	9	174	1.44	25	7	160	3.57
15	11	157	1.36	36	6	45	6.00
12	14	90	0.86	20	8	178	2.50
15	14	130	1.07	30	8	35	3.75
15	11	117	1.36	21	6	2	3.50
20	18	118	1.11	28	6	19	4.67
15	15	143	1.00	31	6	10	5.17
14	13	125	1.08	45	15	43	3.00
16	10	170	1.60	26	6	17	4.33
23	16	114	1.44				
24	20	177	1.20				
AVERAGE = 142/VERTICAL				AVERAGE = 020/80NW			

APPENDIX A

LOC.1554				LOC.1554			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
8	7	100	1.14	24	6	142	4.00
7	5	116	1.40	23	7	138	3.29
8	7	103	1.14	21	7	146	3.00
11	10	109	1.10	40	15	141	2.67
6	5	106	1.20	24	6	166	4.00
18	15	105	1.20	21	5	171	4.20
9	7	86	1.29	18	7	148	2.57
6	5	80	1.20	45	12	146	3.75
7	5	100	1.40	30	10	154	3.00
10	9	80	1.11	25	8	130	3.13
9	7	105	1.29	25	6	157	4.17
				30	7	140	4.29
				28	6	151	4.67
				28	7	140	4.00
				30	8	129	3.75
AVERAGE = 105/70NNE				AVERAGE = 149/68WSW			

LOC.1716				LOC.1716			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
12	11	84	1.09	21	5	117	4.20
14	10	60	1.40	21	4	174	5.25
18	15	74	1.20	27	6	117	4.50
25	18	80	1.39	26	4	165	6.50
16	11	62	1.45	39	8	132	4.88
13	12	74	1.08	29	5	142	5.80
13	11	70	1.18	25	5	146	5.00
12	10	80	1.20	35	10	112	3.50
14	10	70	1.40	40	10	123	4.00
20	14	60	1.43	23	5	116	4.60
19	12	104	1.58	30	6	120	5.00
20	15	95	1.33	23	7	80	3.29
21	14	110	1.50	40	10	118	4.00
13	10	70	1.30	20	5	140	4.00
13	10	65	1.30	25	5	126	5.00
AVERAGE = 085/VERTICAL				AVERAGE = 120/VERTICAL			

APPENDIX A

LOC.1717				LOC.1717			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
11	9	51	1.22	25	6	110	4.17
15	12	2	1.25	32	6	105	5.33
17	13	40	1.31	29	6	91	4.83
8	7	29	1.14	32	6	98	5.33
14	12	38	1.17	24	6	54	4.00
13	10	55	1.30	22	6	110	3.67
10	10	66	1.00	20	5	98	4.00
11	10	42	1.10	22	4	127	5.50
11	8	60	1.38	22	5	92	4.40
13	10	35	1.30	24	6	98	4.00
11	8	52	1.38				
12	9	11	1.33				
16	8	40	2.00				
18	11	60	1.64				
14	10	163	1.40				
15	12	3	1.25				
AVERAGE = 050/VERTICAL				AVERAGE = 095/VERTICAL			

LOC.1718				LOC.1718			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
12	10	132	1.20	31	6	133	5.17
14	9	110	1.56	28	4	175	7.00
13	11	103	1.18	28	6	141	4.67
11	10	102	1.10	18	4	147	4.50
11	10	110	1.10	24	8	135	3.00
8	8	74	1.00	28	7	157	4.00
12	10	130	1.20	28	6	143	4.67
10	9.5	115	1.05	25	5	157	5.00
10	9.5	106	1.05	29	5	169	5.80
12	11	105	1.09	22	5	156	4.40
15	12	50	1.25	24	6	126	4.00
10	9	87	1.11	25	5	149	5.00
9	8	120	1.13	18	4	160	4.50
AVERAGE = 100/VERTICAL				AVERAGE = 150/VERTICAL			

APPENDIX A

LOC.1719				LOC.1719			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
15	13	10	1.15	26	6	20	4.33
20	12	125	1.67	25	6	4	4.17
20	12	147	1.67	25	6	17	4.17
12	11	157	1.09	28	6	38	4.67
12	10	150	1.20	29	6	40	4.83
12	10	148	1.20	23	5	11	4.60
10	8	140	1.25	28	5	17	5.60
11	9	126	1.22	28	7	24	4.00
10	8	144	1.25	30	7	35	4.29
8	7	160	1.14	28	6	5	4.67
6	6	8	1.00	15	5	47	3.00
20	16	163	1.25	25	10	60	2.50
10	10	169	1.00	20	6	20	3.33
17	12	120	1.42	29	6	36	4.83
13	10	174	1.30	25	5	19	5.00
AVERAGE = 145/VERTICAL				AVERAGE = 020/VERTICAL			

LOC.1720				LOC.1720			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
15	12	130	1.25	38	7	54	5.43
9	9	141	1.00	32	7	11	4.57
10	7	158	1.43	52	10	32	5.20
20	11	147	1.82	28	7	67	4.00
16	8	157	2.00	31	8	88	3.88
12	7	3	1.71	32	7	55	4.57
12	6	177	2.00	29	6	22	4.83
12	10	108	1.20	32	8	34	4.00
10	8	117	1.25	29	7	20	4.14
14	7	157	2.00	30	6	40	5.00
12	6	137	2.00	35	10	23	3.50
12	11	20	1.09	31	7	45	4.43
				30	7	30	4.29
AVERAGE = 145/VERTICAL				AVERAGE = 045/VERTICAL			

APPENDIX A

LOC.1744				LOC.1744			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
14	10	13	1.40	18	3	133	6.00
13	7	28	1.86	20	4	120	5.00
20	10	77	2.00	19	4	102	4.75
25	10	66	2.50	20	5	85	4.00
25	10	70	2.50	10	2	130	5.00
10	7	25	1.43	20	4	127	5.00
15	7	41	2.14	17	4	104	4.25
20	10	49	2.00	16	4	110	4.00
25	10	77	2.50	12	3	110	4.00
25	12	55	2.08	20	5	108	4.00
15	10	23	1.50	20	5	82	4.00
7	6.5	30	1.08	20	5	93	4.00
14	7	55	2.00	22	4	91	5.50
15	7	43	2.14	16	4	95	4.00
				20	4	84	5.00
				15	4	113	3.75
				16	4	105	4.00
				15	3	108	5.00
				18	3	86	6.00
				15	3	92	5.00
AVERAGE = 040/79SE				AVERAGE = 090/78N			

LOC.1752				LOC.1752			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
16	8	81	2.00	24	8	110	3.00
12	6	96	2.00	22	4	156	5.50
16	8	134	2.00	18	5	128	3.60
16	8	100	2.00	16	4	162	4.00
14	7	61	2.00	30	5	100	6.00
18	9	176	2.00	23	6	172	3.83
14	7	69	2.00	20	4	176	5.00
30	12	2	2.50	22	5	164	4.40
				21	7	10	3.00
				15	5	150	3.00
				18	4	167	4.50
				18	6	10	3.00
AVERAGE = 080/80SSE				AVERAGE = 160/76WSW			

APPENDIX A

LOC.1762				LOC.1762			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
16	12	149	1.33	30	6	173	5.00
16	8	137	2.00	27	5	25	5.40
17	7	164	2.43	28	6	169	4.67
11	8	168	1.38	36	6	15	6.00
22	8	11	2.75	27	6	2	4.50
18	12	143	1.50	30	6	12	5.00
14	7	137	2.00	21	4	17	5.25
20	10	119	2.00	25	5	12	5.00
15	7	122	2.14	30	5	35	6.00
10	10	120	1.00	27	4	172	6.75
AVERAGE = 130/72NE				AVERAGE = 005/VERTICAL			

LOC.1770				LOC.1770			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
15	10	18	1.50	25	5	13	5.00
20	10	172	2.00	25	5	15	5.00
16	8	157	2.00	20	5	18	4.00
20	10	150	2.00	35	7	15	5.00
20	10	163	2.00	22	4	22	5.50
12	8	170	1.50	20	5	12	4.00
10	9.5	170	1.05	20	5	13	4.00
15	8	175	1.88	25	5	16	5.00
12	8	173	1.50	25	4	20	6.25
12	6	167	2.00	25	5	10	5.00
				30	5	20	6.00
				30	6	16	5.00
AVERAGE = 170/75SW				AVERAGE = 015/82ESE			

APPENDIX A

LOC.1791				LOC.1791			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
18	10	14	1.80	25	5	10	5.00
14	7	175	2.00	25	4	38	6.25
15	8	155	1.88	25	5	40	5.00
20	10	141	2.00	20	3	49	6.67
6	5.75	10	1.04	25	5	12	5.00
12	8	138	1.50	25	5	20	5.00
20	10	166	2.00	20	6	62	3.33
10	9.5	158	1.05	30	5	40	6.00
18	9	147	2.00	25	5	171	5.00
18	9	139	2.00	25	5	43	5.00
16	12	173	1.33				
12	6	145	2.00				
AVERAGE = 155/76ENE				AVERAGE = 035/VERTICAL			

LOC.1804				LOC.1804			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
12	10	82	1.20	18	3	126	6.00
14	13	62	1.08	25	7	136	3.57
19	18	40	1.06	28	4	132	7.00
11	10	64	1.10	24	6	116	4.00
15	14.5	55	1.03	17	4	135	4.25
18	11	21	1.64	24	10	157	2.40
22	15	80	1.47	30	6	143	5.00
14	10	72	1.40	35	8	115	4.38
8	7	68	1.14	30	10	140	3.00
15	12	50	1.25	25	10	148	2.50
20	10	39	2.00	25	6	127	4.17
25	15	10	1.67	23	10	151	2.30
20	10	94	2.00	28	7	117	4.00
15	10	78	1.50	15	3	125	5.00
10	9.5	88	1.05	20	4	128	5.00
10	8	67	1.25	40	10	118	4.00
				16	4	119	4.00
				18	5	126	3.60
				12	4	130	3.00
				24	6	110	4.00
AVERAGE = 065/70NNW				AVERAGE = 130/56NE			

APPENDIX A

LOC. 1886				
APPROXIMATELY UNIMODAL FABRIC				
L.A.(mm)	S.A.(mm)	TREND	RATIO	
28	5	56	5.60	
21	6	23	3.50	
28	5	80	5.60	
26	5	58	5.20	
23	6	39	3.83	
24	6	82	4.00	TILED
20	6	93	3.33	*
15	9	15	1.67	
22	5	44	4.40	
20	5	61	4.00	
25	4	38	6.25	
27	7	32	3.86	
23	6	51	3.83	
29	5	65	5.80	
28	5	44	5.60	
24	4	85	6.00	
12	6	174	2.00	
13	6	86	2.17	
21	4	57	5.25	
18	4	80	4.50	
28	6	88	4.67	
AVERAGE = 058/79SE				
STRONG ALIGNMENT OF ENCLAVES = c. 050				

APPENDIX A

LOC.1916				
APPROXIMATELY UNIMODAL FABRIC				
L.A.(mm)	S.A.(mm)	TREND	RATIO	
21	4	178	5.25	
19	5	4	3.80	TILED
14	9	71	1.56	
26	5	177	5.20	
18	4	175	4.50	
14	4	33	3.50	
22	5	26	4.40	
16	4	30	4.00	
21	4	10	5.25	
24	5	12	4.80	
17	4	170	4.25	
19	4	9	4.75	
22	7	18	3.14	
30	8	119	3.75	
21	3	23	7.00	
20	5	21	4.00	
15	9	40	1.67	
25	5	14	5.00	
14	6	4	2.33	
30	7	163	4.29	
AVERAGE = 013/VERTICAL				

LOC.1917				
AVERAGE FELDSPAR ALIGNMENT				
L.A.(mm)	S.A.(mm)	TREND	RATIO	
23	6	97	3.83	
26	7	91	3.71	
16	6	96	2.67	
21	9	64	2.33	
24	6	82	4.00	
23	5	99	4.60	
24	4	118	6.00	
16	4	90	4.00	
26	6	98	4.33	
16	4	114	4.00	
23	6	102	3.83	
14	4	92	3.50	
15	6	83	2.50	
24	5	114	4.80	
30	7	107	4.29	
22	7	103	3.14	
20	6	97	3.33	
20	5	77	4.00	
16	4	112	4.00	
14	7	106	2.00	
20	6	73	3.33	
16	4	111	4.00	
29	7	115	4.14	
20	7	119	2.86	
20	7	101	2.86	
20	5	77	4.00	
19	5	115	3.80	
19	7	103	2.71	
20	10	109	2.00	
20	5	90	4.00	
25	7	106	3.57	
16	4	103	4.00	
AVERAGE = 110/81NNE				
AVERAGE MAFIC = 095/73N (not measurable individually)				

APPENDIX A

LOC.1918				
AVERAGE FELDSPAR ALIGNMENT				
L.A.:(mm)	S.A.(mm)	TREND	RATIO	
23	5	87	4.60	
20	5	79	4.00	
15	6	68	2.50	
32	11	106	2.91	
13	4	90	3.25	
25	5	76	5.00	
16	8	171	2.00	
21	5	57	4.20	
17	7	86	2.43	
22	4	69	5.50	
20	4	99	5.00	
20	5	74	4.00	
18	5	76	3.60	
21	4	67	5.25	
14	6	62	2.33	
24	5	76	4.80	
22	4	70	5.50	
30	6	95	5.00	
19	4	81	4.75	
22	9	116	2.44	
13	7	80	1.86	
15	5	87	3.00	
22	4	82	5.50	
22	4	92	5.50	
25	5	86	5.00	
22	4	83	5.50	
16	3	90	5.33	
38	7	140	5.43	
29	5	121	5.80	
25	10	76	2.50	
AVERAGE = 088/79S				
AVERAGE MAFC = 077/VERTICAL (not individually measurable)				

APPENDIX A

LOC. 1931				LOC. 1931			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
22	7	62	3.14	26	6	124	4.33 *
16	8	35	2.00	20	4	108	5.00
9	8	47	1.13	22	5	97	4.40
11	7	41	1.57	27	6	63	4.50
12	10	23	1.20	31	5	90	6.20
14	9	55	1.56	23	5	111	4.60 *
16	8	55	2.00	21	4	98	5.25
10	7	58	1.43	26	8	51	3.25
11	8	49	1.38	21	6	74	3.50
17	8	59	2.13	22	4	106	5.50
10.5	10	29	1.05	16	3	80	5.33
14	11	51	1.27	27	5	96	5.40
14	8	39	1.75	36	5	123	7.20 *
11	7	64	1.57	19	8	36	2.38
15	8	53	1.88	20	6	67	3.33
13	8	63	1.63	20	5	77	4.00
9	6	40	1.50	26	7	70	3.71
15	6	72	2.50	26	5	97	5.20
13	10	67	1.30	27	7	142	3.86 *
11	9	51	1.22	24	4	124	6.00 *
				20	5	92	4.00
				31	8	85	3.88
				31	7	93	4.43
				40	8	96	5.00
				37	8	92	4.63
AVERAGE = 048/71SE				AVERAGE = 085/78N			
				* = TOUCHING/TILED			

APPENDIX A

LOC.1932			
APPROXIMATELY UNIMODAL FABRIC			
L.A.(mm)	S.A.(mm)	TREND	RATIO
27	8	94	3.38
33	5	132	6.60
20	5	107	4.00
30	9	91	3.33
25	6	77	4.17
20	5	97	4.00
21	5	82	4.20
22	8	86	2.75
21	8	90	2.63
31	8	110	3.88
11	9	35	1.22
13	7	84	1.86
15	10	67	1.50
18	4	78	4.50
20	9	83	2.22
22	8	88	2.75
32	8	112	4.00
26	9	78	2.89
16	4	99	4.00
33	6	105	5.50
27	5	89	5.40
14	7	77	2.00
32	8	90	4.00
30	7	80	4.29
24	4	95	6.00
22	5	100	4.40
18	5	93	3.60
30	7	111	4.29
AVERAGE = 095/84S			

APPENDIX A

LOC.1934				LOC.1934			
SUBFABRIC A: LOW A.R.				SUBFABRIC B: HIGH A.R.			
L.A.(mm)	S.A.(mm)	TREND	RATIO	L.A.(mm)	S.A.(mm)	TREND	RATIO
10	6	179	1.67	38	7	36	5.43
15	8	4	1.88	50	6	47	8.33
12	8	177	1.50	25	5	41	5.00
12	8	6	1.50	26	5	44	5.20
15	8	20	1.88	31	6	47	5.17
11	7	48	1.57	32	6	45	5.33
10	7	3	1.43	48	6	50	8.00
13	7	13	1.86	40	6	63	6.67
15	11	161	1.36	31	6	48	5.17
15	8	19	1.88	32	5	54	6.40
13	7	146	1.86*	32	7	66	4.57
11	10	173	1.10*	23	5	39	4.60
15	8	19	1.88	20	5	30	4.00
13	7	168	1.86	31	6	48	5.17
12	8	5	1.50	42	5	38	8.40
11	7	10	1.57	28	5	40	5.60
13	6	1	2.17	26	5	51	5.20
20	10	10	2.00	31	5	38	6.20
13	8	25	1.63	20	5	35	4.00
13	9	13	1.44	27	5	43	5.40
10	5	10	2.00	21	4	31	5.25
11	7	166	1.57	40	5	23	8.00
12	6	92	2.00	37	6	24	6.17
				30	4	59	7.50
				30	5	51	6.00
				40	6	60	6.67
				30	5	47	6.00
AVERAGE = 010/83W				AVERAGE = 040/87SE			

APPENDIX A


LOC.1941 SUBFABRIC A; LOW A.R.				LOC.1941 SUBFABRIC B; HIGH A.R.			
L.A.mm	S.A.mm	TREND	RATIO	L.A.mm	S.A.mm	TREND	RATIO
14	9	130	1.56	25	7	23	3.57
13	9	120	1.44	17	6	3	2.83
10	8	126	1.25	29	5	150	5.80
11	8	134	1.38	30	8	167	3.75
12	8	95	1.50	21	3	19	7.00
15	7	117	2.14	28	7	9	4.00
10	7	90	1.43	32	7	50	4.57
11	7	141	1.57	26	8	135	3.25
18	10	109	1.80	20	5	160	4.00
11	10	91	1.10	25	6	154	4.17
11	10	114	1.10	25	5	138	5.00
13	7	105	1.86	32	5	168	6.40
				23	5	4	4.60
				33	9	30	3.67
AVERAGE = 135/81SW				AVERAGE = 012/85E			


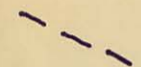


APPENDIX B








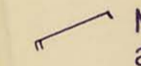



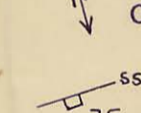
LOCALITY	X	Z	Y	Rxz	
31	230mm	70mm		3.28	1.24
58	40mm	5mm		8.00	2.15
66	70mm	12mm		5.83	1.97
	25mm	9mm		2.77	1.15
	64mm	14mm		4.57	1.65
	154mm	70mm		2.20	0.80
97	90mm	35mm		2.57	0.98
	120mm	50mm		2.40	0.89
155	240mm	70mm		3.43	1.34
161	170mm	35mm		4.86	1.75
	80mm	15mm		5.33	1.87
	60mm	5mm		12.00	2.50
172	160mm	25mm		6.40	1.96
	100mm	20mm		5.00	1.78
	70mm	30mm		2.33	0.87
	100mm	15mm		6.66	2.00
	360mm	80mm		4.50	1.62
178	320mm	60mm		5.33	1.87
	180mm	80mm		2.25	0.82
	90mm	15mm		6.00	1.90
	70mm	20mm		3.50	1.31
194	60mm	25mm		2.40	0.89
204	100mm	15mm		6.66	2.00
	75mm	7mm		10.71	2.40
	75mm	10mm		7.50	2.10
	40mm	5mm		8.00	2.15
333	20mm	10mm		2.00	0.70
341	23mm	19mm		1.21	0.20
421	200mm	30mm		6.66	2.00
		40mm	100mm		
461	180mm	30mm		6.00	1.90
	160mm	25mm		6.40	1.96
	50mm	20mm		2.50	9.40
463	250mm	10mm	160mm	25.00	>2.7
	240mm	10mm	80mm	24.00	>2.7
		240mm	15mm	140mm	>2.5
464	60mm	20mm		3.00	1.14
484	100mm	15mm		6.66	2.00
	75mm	10mm		7.50	2.10
	40mm	5mm		8.00	2.15
543	240mm	25mm		9.60	2.25
550	120mm	30mm		4.00	1.50
	100mm	35mm		2.86	1.10
1600	150mm	55mm		2.73	1.02
	122mm	80mm		1.53	0.42
1721	80mm	42mm		1.90	0.65
	75mm	37mm		2.03	0.61
1804	155mm	57mm		2.72	1.02
	50mm	15mm		3.33	1.27
	70mm	28mm		2.50	0.94
1887	134mm	53mm		2.53	0.92
	86mm	52mm		1.65	0.50
	84mm	49mm		1.71	0.53
	77mm	19mm		4.05	1.53





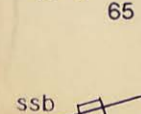


MAP 1

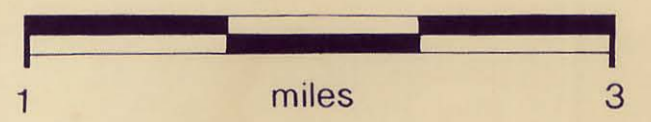
-  Thorr Granite
-  Falcarragh Pelite
-  Ards Quartzite
-  Zone of most intensely mobilised country rocks with prolific granite-filled sinistral shear bands

-  Lithological boundary (certain)
-  Lithological boundary (uncertain)
-  Tectonic slide
-  Pelitic country rock raft
- L.A.S. Lough Anure Sheet

-  Bedding (right way up)
-  Bedding (inverted)
-  General foliation
-  Mineral lineation
-  Crenulation cleavage
-  Early fold axial plane with strong axial planar cleavage
-  Early fold plunge (vergence indicated)
-  Main fold axial plane at Lough Agher
-  Main fold plunge at Lough Agher (vergence indicated)
-  Main fold axial planes in mobilised country rocks
-  Main fold plunge in mobilised country rocks (vergence indicated)
-  Inclined sinistral shear band

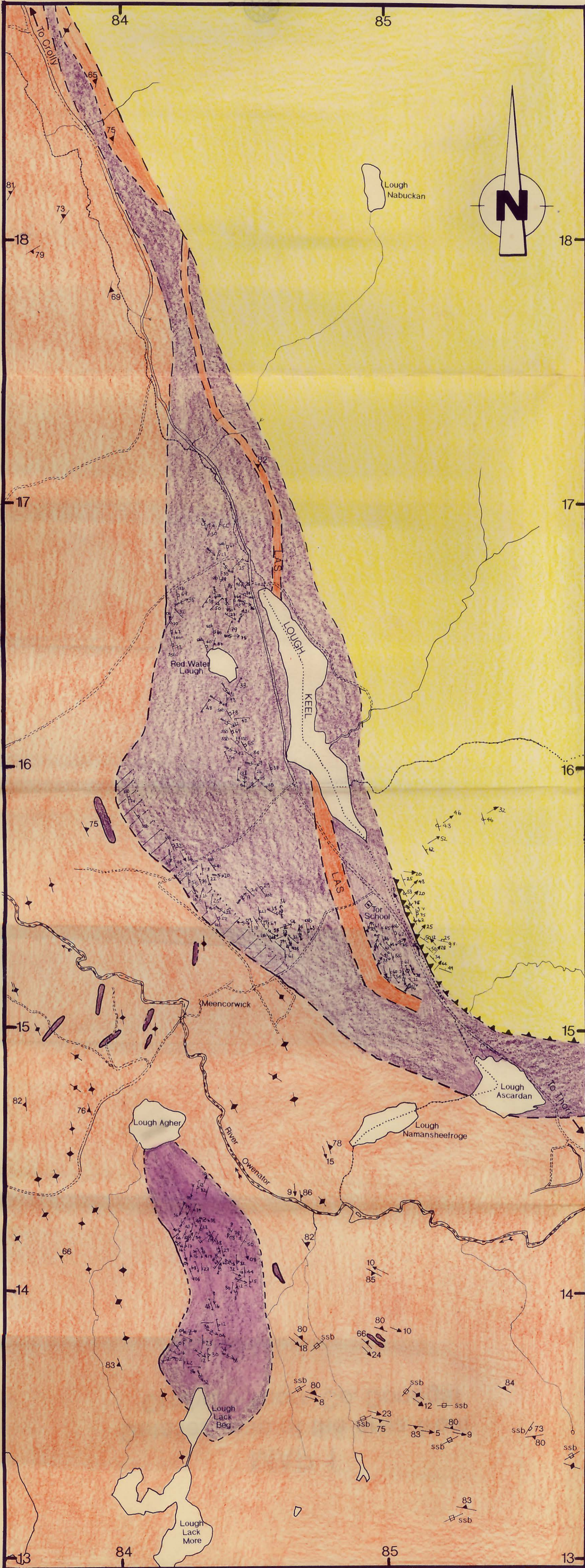
-  Inclined foliation in Thorr Granite
-  Vertical foliation in Thorr Granite
-  Mineral lineation in Thorr Granite
-  Inclined sinistral shear band
-  Vertical sinistral shear band

Scale = 1 : 10 560



(approx. 1" = 1 mile)

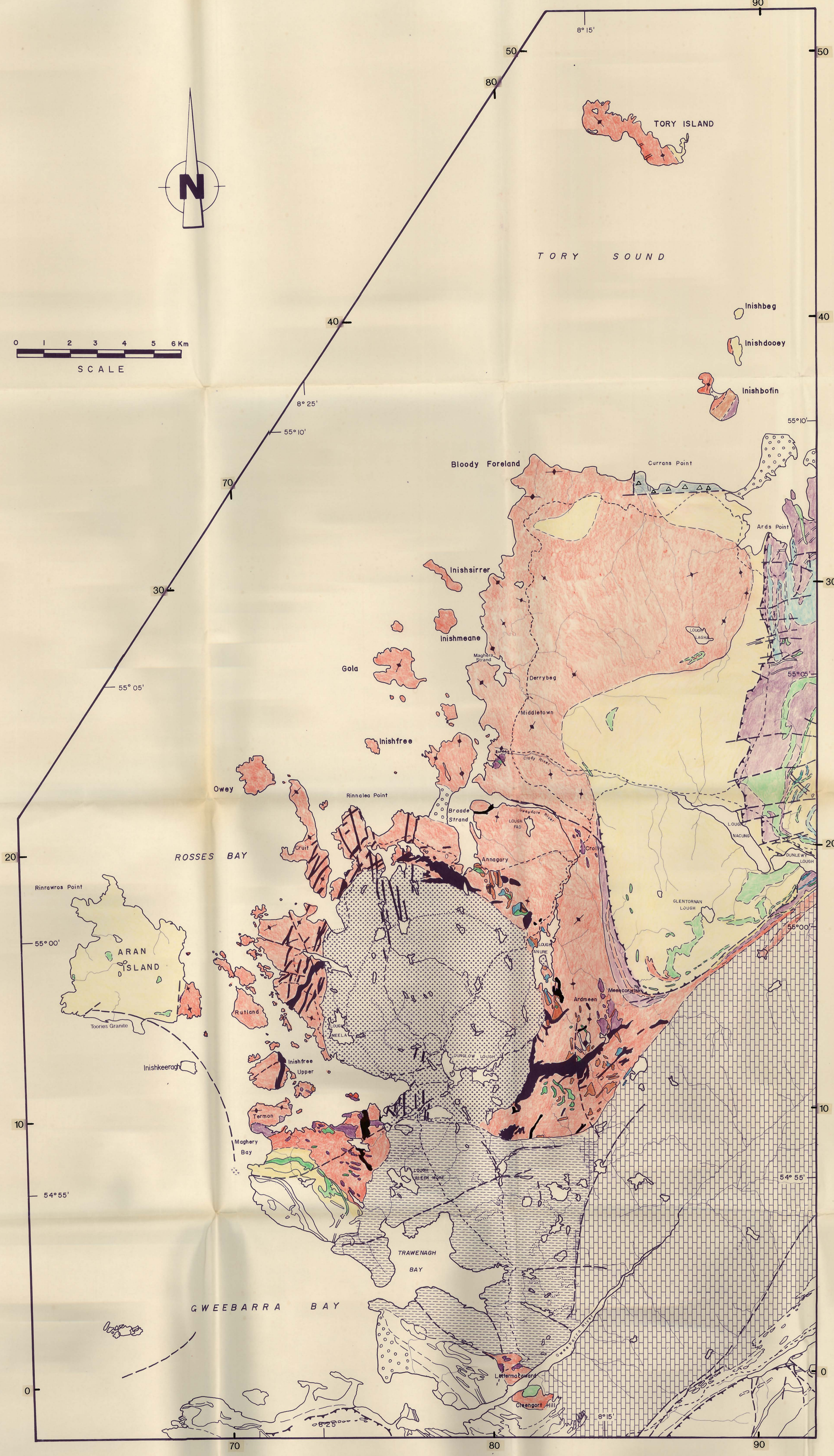
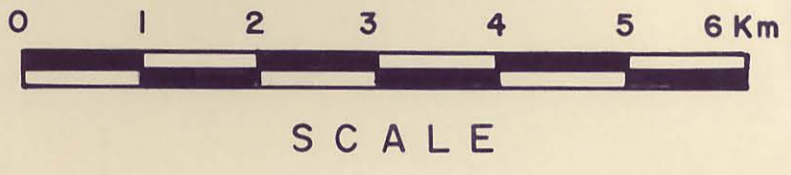
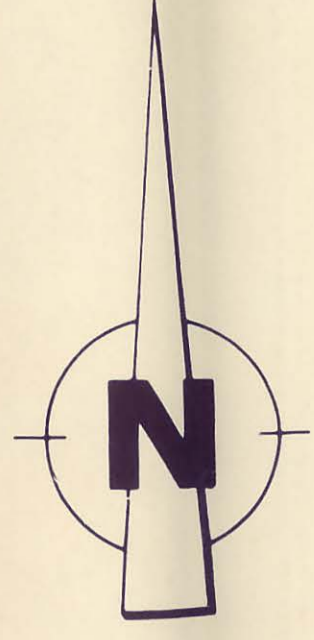
Grid transposed from 1 : 25 000 scale





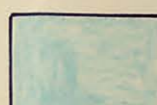
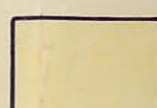







MAP 2

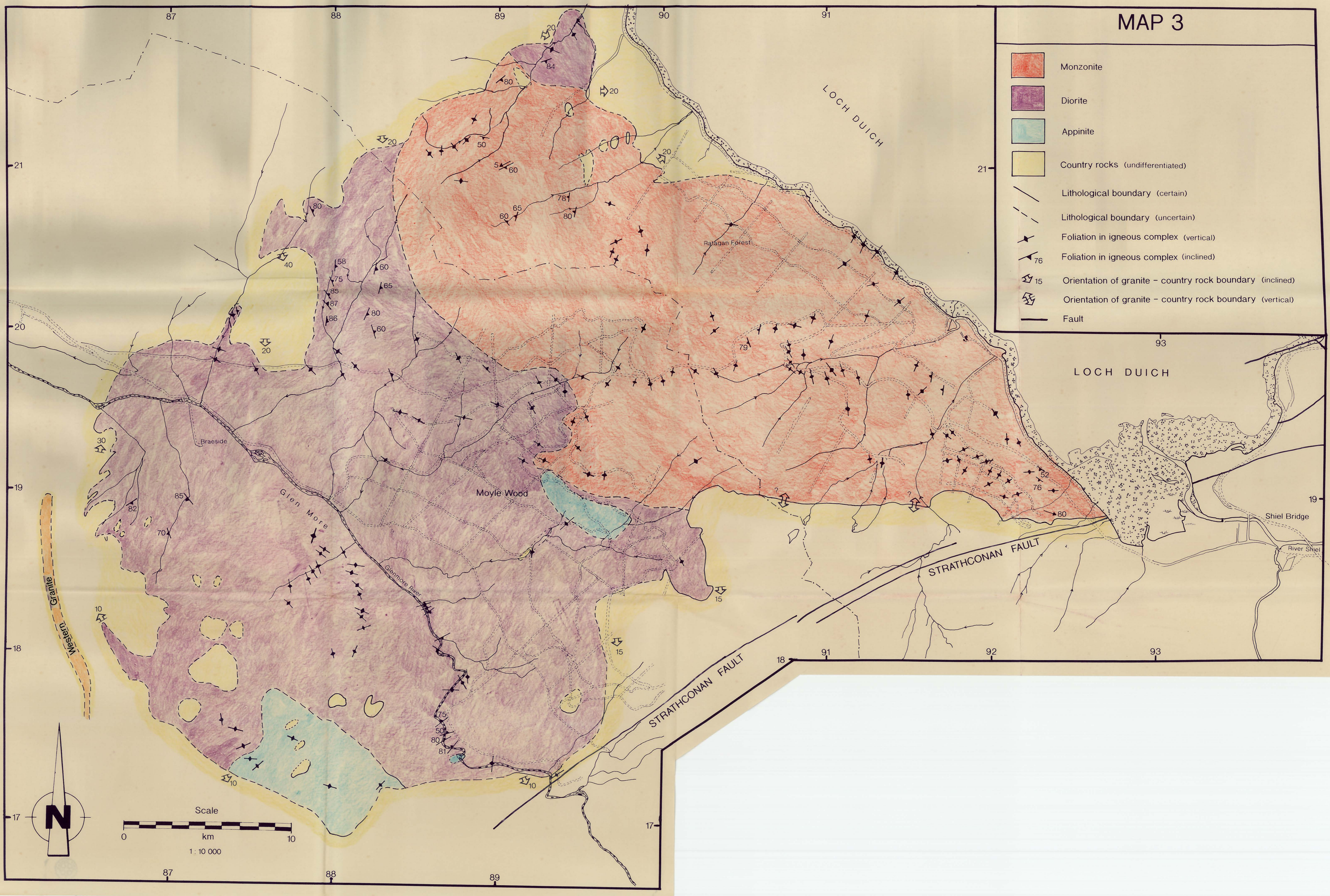
LEGEND

- Lithological contact (certain)
- Lithological contact (uncertain)
- Tectonic slide
- Fault
- Strike-dip of foliation in granite (vertical)
- Strike-dip of foliation in granite (inclined)
- Quartzite
- Pelite
- Semi-Pelite
- Thorr Granite
- Rosses Granite
- Main Donegal Granite
- Appinite
- Limestone
- Metadolomite
- Travenagh Bay Granite
- Aplite
- Road



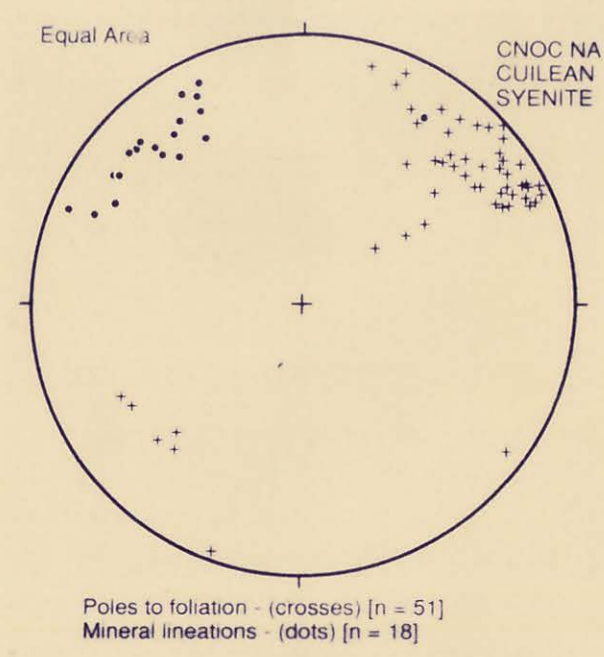
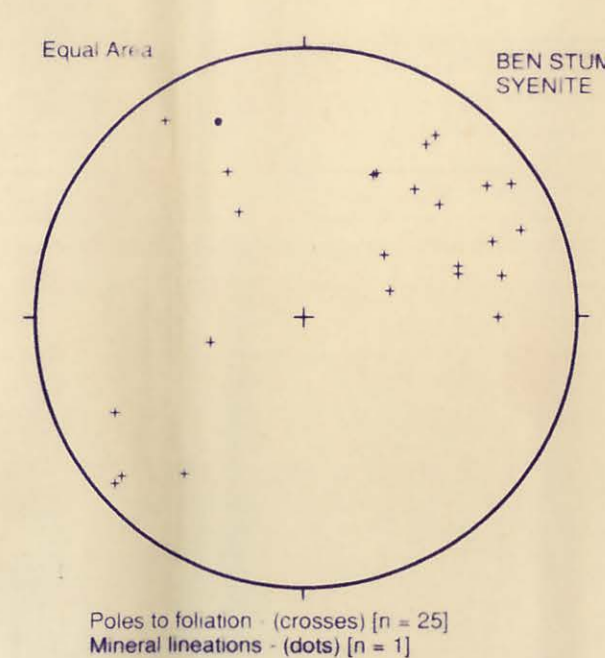
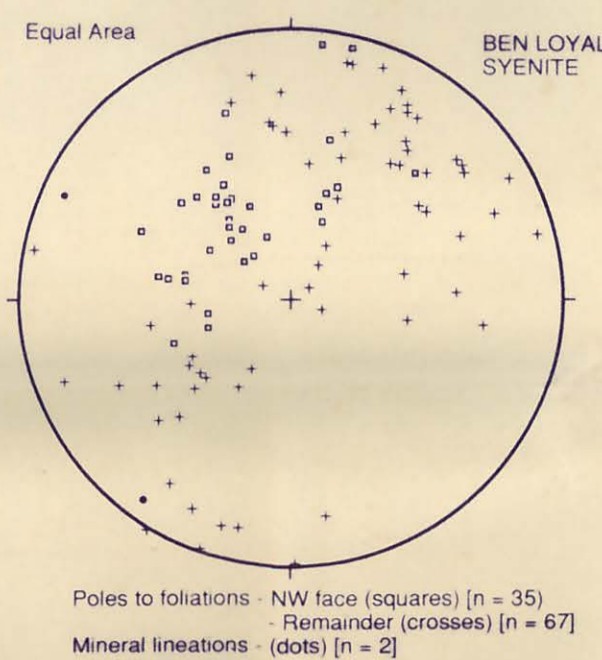
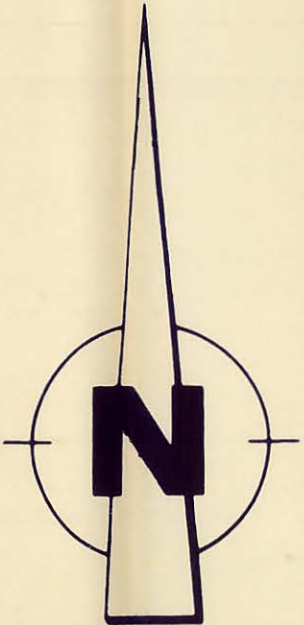
MAP 3

	Monzonite
	Diorite
	Appinite
	Country rocks (undifferentiated)
	Lithological boundary (certain)
	Lithological boundary (uncertain)
	Foliation in igneous complex (vertical)
	Foliation in igneous complex (inclined)
	Orientation of granite - country rock boundary (inclined)
	Orientation of granite - country rock boundary (vertical)
	Fault





0 Km 1
Scale = 1 : 10 000



MAP 4

- Ben Loyal Syenite (outer facies)
- Ben Loyal Syenite (core facies)
- Ben Stumanadh Syenite
- Cnoc nan Cùilean Syenite
- Country rocks (undifferentiated)
- Fault
- Cross folds
-
- = 0 - 30°
- = 30 - 60°
- = 60 - 90°

Orientation of foliation in Syenites

