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THE EMPLACEMENT AND DEFORMATION OF GRANITIC ROCKS IN A TRANSPRESSIONAL SHEAR ZONE: THE OX MOUNTAINS IGNEOUS COMPLEX

by

Kenneth J.W. McCaffrey

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A thesis submitted for the degree of Doctor of Philosophy at the Department of Geological Sciences, University of Durham.

December 1989.
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ABSTRACT

The structural evolution of the Ox Mountains Granodiorite (478±12Ma) during and after its emplacement is described. This pluton has been emplaced within and synchronously with, a major transpressional shear zone which is expressed as a 11km wide belt of strongly deformed NE-SW striking, steeply dipping metasediments. The steep shear zone cleavage is intensified to a mylonitic fabric in a braided system of high strain zones formed throughout the Ox Mountains Inlier. Pervasive ductile, sinistral deformation in the shear zone is interrupted by the emplacement of the pluton and by the synchronous development in the country rocks of series of brittle thrust structures, which produced a displacement upwards and towards the centre of the shear zone. These thrusts are intimately associated with the emplacement of moderately inclined granodiorite sheets belonging to the main intrusive phase indicating a component of vertical extension in the country rocks at this time.

The OMG is a heterogeneous, four component pluton internally composed of a series of large sheets or dykes. Minor muscovite granite sheets emplaced along the northern contacts, preceded the main intrusive sheets of Group 1 and Group 2 granodiorite with associated diorites. Sheets of tonalites and minor components completed the emplacement history. A prolonged history of sinistral transpressional shearing has deformed the pluton. Discrete sinistral shears indicate an early localization event is overprinted by a main ductile penetrative fabric which cross-cuts all internal contacts. This foliation is deformed by extensive sinistral S-C fabrics. Later deformation becomes increasingly partitioned into late sinistral and dextral shear zones which are locally mylonitic. Microstructural evidence suggests that the main foliation was formed under lower amphibolite facies and deformed by a steady state flow process. The S-C fabric and late shear zone formation, best developed in the granodiorites and granites, may have been initiated by a switch in the predominant alkali feldspar deformation mechanism from crystal plastic to a diffusive mass transfer process. This may be a retrogressive effect, and the product is a grain size reduction which may lead to ultramyolite production.

The emplacement model for the Ox Mountains Granodiorite is constrained by the original geometry of the dykes or sheets. These data rule out emplacement of the OMG in a releasing bend or pull-apart structure. Strain data does not allow a forceful mechanism and a permissive emplacement model is preferred, in which vertical extension during the intrusive episode created an area of dilation in which dyking occurred. This was caused by oblique movements on two upwardly converging high strain zones outside the pluton. Two satellite plutons, the Lough Talt Adamellite and the Easkey Lough Adamellite were emplaced in extensional cavities created by reactivated sinistral movements on one of the high strain zones at a much later date. (c400Ma).

Transpressional shear zones may initiate or enhance melting in the lower crust and mantle lithosphere where thermal perturbation has occurred. The shear zones may provide conduits for the melts and emplacement sites, especially where high transpressional strains are accommodated by vertical uplift.
Acknowledgements

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References.  

A geological map of The Ox Mountains Inlier and a lithological and structural map of the Ox Mountains Granodiorite is included at the back of this thesis which is intended for use with Chapters 2 to 5 and Chapter 7.
...the interpretation of field work is often a strictly personal affair, depending on the observer's character, training and experience. Field experience is often highly specialized and thus may tend to biased general opinions; in any event it will differ greatly among different field geologists. From all these almost accidental circumstances, geology as earth history has always been enlivened by debates and controversies, often acrimonius and always vehement.

CHAPTER 1

GRANITOID EMPLACEMENT, DEFORMATION

AND THE OX MOUNTAINS

1.1 Introduction

Most granitic rocks are associated with zones of plate tectonic activity which can be related to deformation at continental margins. These zones include most of the common plate tectonic settings, e.g. subduction, continental collision, transcurrent and extensional type zones. Subduction zones contain the largest volume of granitic material although these are not strictly granitic but tonalite, diorite and granodiorite in composition. The most spectacular examples are located along the western margin of the N and S American continents in the large Cordilleran batholiths (Pitcher 1978, Silver and Chappell 1988). Continental collisional zones contain the true granites, e.g. the High Himalayan leucogranites and Hercynian granites and much of these are derived from anatexis of initially sedimentary material. Large inter-continental transcurrent fault zones also contain granites which may be localized in dilational zones (Sylvester 1988), e.g. the late Caledonian granites of the British Isles. Extensional regimes are commonly associated with basic magmatism but, as in the British Tertiary complexes, granites are produced by fractionation of basic magma. A consideration of the diversity of granitic types and environments makes it obvious that the Read (1957) dictum that there are 'granites and granites' is certainly still valid for current ideas on granite production related to plate tectonic theory.
An important division is that between the true granites of the migmatitic continental collisional belts and the vast tonalitic associations of the subduction related granitoids which Read (1957) and Buddington (1959) incorporated in the 'Granite Series'; an hypothesis which related emplacement style to crustal level. High level discordant plutons typical of Cordilleran batholiths were termed 'magma intrusive' or 'epizonal'. 'Autochthonous' or 'catazonal' plutons were formed at much deeper levels and were typified by the gneiss massifs of the migmatitic and metamorphic zones. 'Parautochthonous' or 'mesozonal' occurred at intermediate depths and contained characteristics of both end members. The work of Pitcher and Berger (1972) on the Donegal Batholith, in which many differing emplacement styles exist at the same crustal level, did not substantiate the granite series approach. Pitcher (1978, 1979) considered that granite type and emplacement style could be related to the tectonic regime into which it had been emplaced. The tectonic approach has been developed by Hutton (1988a) who considered that granite emplacement mechanisms are due to the 'the interaction between natural magma buoyancy forces and ambient tectonic forces'. There is a growing realization that granitic magma generation, ascent and emplacement are all fundamentally related to tectonic activity in the continental lithosphere (Karlstrom 1989, Patterson 1989).

Granites can be emplaced as plutons in a wide variety of shapes and sizes and can display a wide range of structural diversity. The range of structures (including fabrics) which are encountered during the field observation of granitic plutons is more likely to reflect the emplacement mechanism than provide any information on the ascent of the magma from its source. The structures preserved are largely a response of the granitic magma to wall rock tectonic forces and internal magma buoyancy. This response is governed by the physical properties of the magma which can determine the ability to record
deformation. Some magma chamber processes such as convectional overturn of magma would tend to destroy that record but the physical properties of the magma (viscosity in this case) also controls the operation of these processes (McBirney & Murase 1984). The physical properties or rheology of granitic magmas can change markedly with crystal percentage, time, strain, strain rate, water content, temperature and pressure and this results in a correspondingly marked change in structures which will be preserved.

This thesis seeks to explore the differing emplacement styles and structures which form during granitoid emplacement in a major shear zone and to link their development to the changing rheological properties of the granite and its wall rocks. The approach is mainly by detailed field based observation of the geometry and nature of the structures in a pluton combined with microstructural evidence and the use of existing experimental data on granitoids so as to produce a model for the evolution of a deforming pluton throughout its crystallization and post-consolidation history.

1.2 Gravitoid Rheology

Rheology can be defined as the science of flow of matter; it is the study of elasticity, viscosity and flow; it is the relationship between stress, strain and strain rate. Materials can exhibit three main types of behaviour when exposed to an external stress field.

(1) Elasticity.

Stress \((\sigma)\) is proportional to strain \((\varepsilon)\) which is recoverable. It can be defined by Hookes' law as :-

\[
\sigma = \varepsilon
\]  

1.1
(2) Yield

Materials flow above a critical stress termed the Yield stress ($\sigma_0$). Below the yield stress the behaviour is usually elastic.

(3) Viscous Flow

The simplest form of viscous flow is Newtonian viscous flow where stress is proportional to strain rate ($\dot{\varepsilon}$)

$$\sigma \propto \dot{\varepsilon}$$

or

$$\tau = \eta \dot{\varepsilon}$$

where $\tau$ is shear stress and $\eta$ is the viscosity constant for Newtonian viscous flow.

Newtonian fluid behaviour is demonstrated by most pure fluids and has been proposed for silicate melts (Shaw 1965). It was Scrope (1872) who introduced the term ‘magma’ stressing its analogy to compound liquids or solid particles suspended as a fluid. Other similar compounds include blood, mud, paste and these behave in a non-Newtonian manner in response to stress.

Granitic magmas demonstrate a range of physical properties during crystallization. Pure granitic melts have been modelled as Newtonian fluids (van der Molen & Paterson 1979) the viscosity of which depends on chemical composition and temperature and is strongly dependent on the volatile content. If the magma contains suspended crystals the viscosity becomes strongly dependent on the crystal to viscous liquid ratio. At low crystal contents (less than 50 %) the behaviour can be still regarded as that of a Newtonian fluid with a viscosity which is dependent on the melt viscosity and the percentage of crystals. It is also dependent on the size, shape and distribution of crystals in the magma (Mc Birney & Murase 1984).
For crystal to magma percentages greater than 50% there has been a limited amount of experimental work which suggests that there is a rapid (non-linear) increase in viscosity (van der Molen & Paterson 1979, Arzi 1978) (Fig. 1.1) with increase in crystal percentages. This occurs as a result of the increasing amount of interaction between grains until an essentially solid behaviour is exhibited. This transition, termed the critical melt percentage (CMP), has been assigned a value of approximately 30-35% by van der Molen & Paterson (1979) who find a broad agreement with the results from soil mechanics, melting experiments in garnet lherzolite, viscosity of dense suspensions and the phenocryst concentrations found in dykes.

Above the critical melt percentage the behaviour will become increasingly dependant on the solid framework and its ability to deform in a plastic manner. The likely deformation mechanism would be a dislocation creep or 'power law creep' mechanism (Tullis & Yund 1980). This involves the intra crystalline movement of defects or dislocations through the crystal lattice.

Experimental deformation of granites with differing crystal percentages is difficult to perform due to the wide range of viscosity (therefore strains) which occur around the CMP. This can be as much as $10^{14}$ orders of magnitude (Wickham 1987) illustrating the dramatic change in properties which may occur in a crystallizing granite. There are still many uncertainties (Mc Birney & Murase 1984) in modelling the exact changes which do occur, e.g. the time dependance of stress and strain and the changing viscosity of the melt as phenocrysts are produced during crystallization.

The other important change in physical properties which can occur is that the granite will behave in a non-Newtonian manner as crystals form in the magma. It will possess a yield strength below which deformation will only take place in an elastic manner. Yield strengths have been measured in Hawaiian lavas by Shaw (1968) but as yet have not been satisfactorily measured for
Critical melt percentage (CMP) according to

- Arzi (1978)
- van der Molen & Paterson (1979)
granitic magmas. There are uncertainties in the importance attached to the 
effects of silica content and the lower temperatures at which granitic magmas 
are formed (Mc Birney & Murase 1984) and the basic question as to whether 
the the yield strength is purely a function of the crystallizing matrix or a 
property of the viscous magma has not yet been fully resolved.

These limited experimental findings have been carried out for granites 
alone and little attention has been paid to other compositions such as tonalite, 
diorite and granodiorite. In order to obtain some information about these rock 
types it is useful to examine their field relations which will provide qualitative 
rheological information.

1.3 The development of granitic structures

It would be unwise to attach too great an emphasis on the application of 
results from experimental studies which have been carried out at geologically 
unrealistic strain rates. In general the rheological models do provide a useful 
framework in which the development of granitic structures can be viewed.

The context in which the term granitic structures is used in this thesis 
refers to the development in a pluton of any of the following: fabrics or fo­
liations, contacts, gneissose foliations, shear sense indicators, mylonite zones, 
br brittle faults and breccia development. A brief description of some of the more 
important structures which can form is presented below.

1.3.1 Magmatic state deformation

At low crystal to viscous liquid percentages, early formed euhedral phe­
nocrysts (usually feldspars and biotites in granitic rocks) and rigid xenoliths 
of country rock or igneous material in the presence of a stress field will be 
rotated into parallelism (Gay 1968). The amount of rotation depends on the 
shape ratios of the phenocrysts or xenoliths and they will not be internally
deformed if the viscosity contrast between themselves and the viscous magma is too high. If deformation ceases at this point the fabric produced will be one in which aligned objects are present in an isotropic matrix of granitic magma. Hutton (1988a) termed this a 'pre-full crystallization fabric' and this is equivalent to the 'magmatic state deformation' of Blumenfeld and Bouchez (1988). Their principle criterion for separating this type of fabric from that formed during post-consolidation deformation was the absence of intracrystalline deformation in quartz. This is taken to be the most ductile phase in most quartzo-feldspathic rocks and is usually formed late in the crystallization history. The use of this criterion is dependant on the assumption that most post-consolidation deformation has occurred at temperatures lower than the solidus and a major recrystallization of quartz has not taken place. It is likely in this situation that evidence for this type of deformation and metamorphism would be recorded in the metamorphic history of the country rocks.

At higher (> 70%) crystal percentages the phenocrysts will begin to interact and a solid framework will form which will be capable of transmitting stress and deforming in a plastic manner. If deformation ceases before complete crystallization, pools of quartz may form at grain boundaries and be preserved in an undeformed state.

1.3.2 Solid state deformation

Deformation of a pluton at or below the solidus temperature will produce strain in all mineral phases of the granitoid and will take place predominantly by a process of dislocation creep. The fabric produced has been called a 'crystal plastic strain' fabric (Hutton 1988a) and is referred to as a 'solid state deformation' by Blumenfeld and Bouchez (1988). The latter term is preferred in this thesis as its use does not require a knowledge of the deformation mechanisms which are operating during the formation of the structures.
1.3.3 The Cloosian paradigm

The classification of granitic rock structures developed by Hans Cloos (Balk 1937) has for many years formed the basis for the study of fabric and structures formed in igneous rocks. Two types of structures were held to occur.

(a) Primary structures.

These were formed during consolidation of the magma by flow parallel to the wallrocks. These were regarded as alignments of minerals in platy or linear flow fabrics. Joints were regarded as primary structures as these often bore a geometrical relationship to the flow alignments and were regarded as a continuation of the elongation of the magma.

(b) Secondary Structures.

These were formed during sub-solidus deformation and were thought to be essentially metamorphic. The production of foliations which cross cut internal contacts and the formation of boudins were regarded as examples of this type of deformation.

This philosophy has been largely followed by Marre (1986) and has been criticized in the work of Berger and Pitcher (1970), Pitcher and Berger (1972) and Hutton (1988a). The main concern of latter authors was the ability of the magma to flow in the sense that Cloos and Balk had implied. Their primary structures were considered to be formed by 'magmatic currents' flowing against the wall rocks almost directly analogous to water flowing in a stream. Silicate melts are considered not to behave in this manner but as viscous magmas (see section 1.2) and can be treated in structural terms as a normal geological material.

Berger and Pitcher (1970) questioned the validity of using joints to constrain granite tectonic models and pointed to the difficulty in distinguishing primary jointing (during consolidation) and secondary jointing which has been
ascribed to regional stresses and uplift. They did however admit that joint systems could form very early in the consolidation history of plutonic rocks from evidence of dyking and veining but note that early formed joints that do not contain dyke material are not likely to be preserved after late and post-consolidation processes had modified the early formed igneous textures.

1.3.4 Strain analysis

The magmatic state or solid state fabrics may be classified as either planar or linear or a combination of both elements and can be qualitatively related to the strain ellipsoid in the Flinn (1965) nomenclature (Fig. 1.2). The strain ellipsoid records the relative amounts of deformation in three orthogonal directions of an initially spherical portion of the granite. S-type planar fabric record oblate or flattening strains, L-type linear fabrics record constrictional 'pencil' type strain with the more common LS-type being intermediate between these end members.

In many plutons, the nature of the fabric can be qualitatively determined by observation of the degree of preferred orientation and elongation of the constitutive minerals. A semi-quantitive estimate of the shape of the strain ellipsoid may be obtained by the measurement of the shape ratios of xenoliths or enclaves contained in the granitoid (Hutton 1982, 1988b). Microdiorite or mafic enclaves are considered to be the most useful as they appear to have had an originally spherical shape. The relative length to width ratios for xenoliths in the plane of the foliation (XY) and perpendicular to the plane of the foliation and parallel to the lineation (XZ) on two surfaces at right angles can lead to an estimation of the overall shape of the strain ellipsoid and can be plotted on a Flinn (1965) plot.

A further useful estimation of the shape of the strain ellipsoid can be obtained by use of the 'Fry Method' (Hanna & Fry 1979, Fry 1979) which makes use of the distribution of phenocrysts. This method can conveniently
The Flinn diagram

K = oo

1+e1

1+e2

K = 1

uniaxial prolate

plane strain

1+e3

1+e2

uniaxial oblate

K = 0

Fig. 1.2

(flatening strain)

(after Flinn 1965)
be used both in the field or in the laboratory.

### 1.3.5 Shear sense

In many of the crustal regimes in which granites can be emplaced the deformation of both wall rocks and granite is non-coaxial. Simple shear is one representative model for this deformation. A notable exception to this is ballooning plutons which produce a predominantly flattening (S-type) fabric and strains. Non-coaxial deformation leads to the development of asymmetries in rock structures which are known as 'shear sense indicators.' These usually occur on a mesoscopic or microscopic scale and can provide information as to the overall kinematics of a crustal deformation. Shear sense indicators must be viewed perpendicular to the main foliation and parallel to the direction of transport or stretching lineation. There are a number of different shear sense indicators which can be used on both magmatic and solid state deformations. Blumenfeld and Bouchez (1988) cite three main shear sense indicators for magmatic state deformation (Fig. 1.3).

1. The obliquity between the deforming walls and a planar foliation formed by early formed phenocrysts will give a sense of shear (Fig. 1.3a).

2. The obliquity between sub-fabrics which is the obliquity of different mineral phases due to their differing rates of rotation whilst undergoing simple shear (Fig. 1.3b).

3. The sense of tiling of megacrysts; the rotation of phenocrysts in a viscous medium can produce an imbrication type structure when they touch. Statistical counting of populations of megacrysts will show a majority in the shear sense direction (Fig. 1.3c).

Solid state shear sense indicators can be applied to both wall rock and granitoid deformation. The application of shear sense indicators has been well established over the past 10 years. It was the publication of Berthe et al. (1979) which first applied S-C indicators to granitic rocks. These have
Magmatic state shear sense indicators

(a) obliquity of magmatic C to S plane

(b) obliquity between sub-fabrics

(c) tiling of megacrysts

(From Blumenfeld & Bouchez 1988)
also been applied to metamorphic rocks by Lister and Snoke (1984). Two sets of planar anisotropies can occur one termed S or schistosite' (schistosity or foliation) and the other C or cisaillement (shear). The C surfaces are initiated and remain parallel to the shear zone boundary during progressive deformation. They are considered to be spaced slip surfaces with a sense of shear which is the same as the major shear zone. The S surfaces are formed at an angle of 45° or less and curve asymptotically into the C surfaces as in Figure 1.4a.

Other shear sense indicators (Simpson and Schmid 1983) include asymmetric augen structures in mylonitic gneiss and asymmetric pressure shadows on porphyroclasts which have a high ductility contrast with the matrix, and are rotated (usually with the same sense of vorticity) as the shear zone (Fig 1.4b).

Porphyroblasts which are growing in a deforming matrix will have inclusion trails which are curved with respect to the external fabric and these can provide useful timing constraints on the relationship between metamorphism and deformation in the wall rocks (Fig 1.4c).

Crystallographic fabrics can be asymmetric with respect to the foliation and lineation. Numerical models have been used by Lister and Hobbs (1980) to simulate the fabrics which may be developed (Fig. 1.4d). Quartz is the most commonly used for shear sense determination and the work of Law et al. (1986) and Gapais & Cobbold (1987) has shown how quartz fabrics can differentiate between flattening and simple shear regimes and can provide an estimation of the shape of the finite strain ellipsoid. Plagioclase fabrics have also been used (Jensen & Starkey 1985) although these are necessarily more complicated due to the biaxial nature of the feldspars. Biotites can also provide kinematic information (Marre 1986) but are rarely investigated.

These shear sense indicators are essentially those which form when a ma-
Solid state shear sense indicators

Fig. 1.4

(a) S-C fabrics

(b) asymmetric pressure shadows

(c) curved inclusion trails

(d) asymmetric quartz fabrics
terial is deforming in a ductile manner. Variations in strain rate, pressure, temperature and water content may cause a combination of brittle and ductile deformation. A common example of this phenomenon is that of feldspars (particularly plagioclase) exhibiting brittle features while coexisting quartz is deforming in a ductile manner. Displaced or broken grains can be shear sense indicators but in many cases the sense of shear on the microfractures will be opposite to that of the overall shear regime (Simpson & and Schmid 1983).

In some cases marker horizons such as banding in granites and internal contacts may be deflected or displaced providing a good indicator of sense of shear as well as providing a value for the displacement distance. Shear sense indicators associated with brittle fault behaviour, although not discussed here include slickenfibres and striations and a variety of secondary fractures (Petit 1987).

This brief discussion provides a summary of the main criteria which are available for the field and laboratory determination of sense of shear. In general they should not be applied in isolation because the existence of zones of antithetic shear are common. Microstructural shear indicators should be reinforced by the kinematic evidence which has been accumulated in the field and were possible by the formation of megascopic structures.

1.3.6 Contacts

Contacts can be generally classified as concordant or discordant. Marre (1986) states that the main factors which determine the nature of a contact are :-

(1) The respective volume of each material
(2) The respective viscosity of each material.

If there are equal quantities of material then the contact nature is determined by the relative viscosities of each. If the viscosities are the same the contact will form a straight line and a foliation will pass straight through both
materials forming a concordant contact. If there is a strong viscosity contrast then an irregular blocky contact may form where a foliation in one material is truncated by the contact with the later material thus forming a discordant contact.

1.4 Emplacement mechanisms

The basic methodology and philosophy regarding granite rheology and structural evolution can be applied in the field examination of a pluton with the aim being to produce detailed maps which will aid the overall development of an emplacement model which will take into account the structural evolution of both the wall rocks and the granitic rocks.

Of particular importance is the timing of the regional or local deformation events relative to the intrusive events. The work of Pitcher and Read (1959) and Pitcher and Berger (1972) was among the first to establish the timing of granitic intrusion with respect to deformation in the aureole. This was modified by the work of Hutton (1977) who established a chronological sequence of deformation and metamorphism throughout N.W. Donegal into which the Donegal plutons were intruded at a late stage. The relative timing of plutonism and deformation may be thoroughly established by the use of modern kinematic studies and pluton age dates, obtained by a variety of isotopic methods, can be used to date deformation events which are occurring in the aureole. It is important to take into account the possible resetting of age dates due to later deformational and metamorphic events (Page & Bell 1985).

The structural maps containing foliation, lineation and contact details and the wall rock structural information may be used to interpret regional structures and intrusion events to produce an emplacement model. Local emplacement models do not necessarily reflect the larger scale emplacement
of a batholith but may still be of use in constraining a larger scale model.

Emplacement mechanisms in general can be responsible for the shape and structural characteristics of a pluton. They can in some cases give information about the ascent mechanism but in others they do not (Pitcher 1979). Ballooning plutons can be fed by magma which ascends through fractures in the lower and middle crust (Bateman 1984). Emplacement has been considered in terms of a forceful or passive type mechanism.

1.4.1 Forceful emplacement

Forceful emplacement implies displacement of crustal rocks to create a volume for granitic magma. These are often regarded as ballooning diapirs and some notable examples include, Ardara (Pitcher & Berger 1972, Holder 1979) Rogart (Soper 1963), Papoose flat (Sylvester et al. 1978), Cannibal Creek (Bateman 1984) and the N. Arran Granite (England 1988). These plutons often show evidence for expansion with intense flattening strains developed both within the pluton (not always) and in the wall rocks. The intense deformation in the pluton around the margin as in Ardara may be due to the multiple injection of magma in the centre as this deforms the colder marginal facies.

1.4.2 Passive emplacement

During passive emplacement crustal rocks will be replaced generally by by the stoping of large blocks to create a space in which magma may be emplaced. A passive stoping mechanism was advocated by Daly (1933) to account for the large discordant cauldrons which form in the Sierra Nevada Batholith. Other examples include the Fanad and Thorr plutons in the Donegal batholith (Pitcher & Berger 1972). A variation is where the country rocks are displaced by tectonic means to create a space into which granitic magma is drawn, e.g. the Strontian biotite granite in Scotland (Hutton 1988b) where an extensional shear zone created a flat space for the granite to be emplaced. The formation
of ring dykes and cauldrons (Clough et al. 1909) is another example of passive emplacement.

1.4.3  Tectonic contols on emplacement

The role of crustal deformation and tectonics in general has in recent years been assuming greater importance for the problem of granite emplacement. Anderson (1936) considered the importance of magma pressure relative to the stress regime surrounding a central complex as the prime factor in controlling the emplacement of ring dykes and cone sheets. Murrell (1970) acknowledged the overwhelming relationship between global tectonic zones and igneous activity by showing the coincidence of seismic zones and volcanic zones with plate margins. The work of Pitcher (1979, 1984 & 1987) has attempted to classify granitoids according to their tectonic setting.

The problem of the emplacement of the Donegal Batholith has provided a forum for continued debate throughout the history of granite research. The debate has mainly centred on the metamorphic versus magmatic origin for the Donegal granites and was especially vociferous during the mid to late 19th century and the 1950s. It was from their work on the Donegal granites that Pitcher and Berger (1972) largely disproved the Granite Series approach of Read and Buddington, by showing that crustal level was not the main factor in controlling the emplacement style of granites. Hutton (1982) produced an integrated model in which the emplacement of seven plutons could be related to the operation of a sinistral shear zone into which the Main Donegal pluton was emplaced. A shear zone can produce elements of compression and extension in tip and bend zones (Sanderson & Marchini 1984) and these can be related to the emplacement styles of the plutons in Donegal. Dilational zones produce large, passively emplaced, plutons which contain large stopped blocks. Compressional zones produce forcefully emplaced ballooning diapirs such as the Ardara and Toories plutons. The main Donegal Granite itself
exhibits a sheeted structure formed by the wedging of magma into the shear zone.

Other examples of shear zones which contain granites are described by Brun and Pons (1981). The foliation patterns which are produced by balloon-ing plutons into transcurrent and thrust geometry shear zones show interference patterns between the regional and balloonning strains that form triple points which, depending on the relative amounts of each process, may be located inside or outside the pluton. The Mortagne pluton (Guineberteau et al. 1987) was emplaced in a dilational regime which was produced by a sinistral shear zone in the S. Armorican shear zone. Davies (1982) has described the emplacement of plutons in Saudi Arabia (the Ajjaj shear zone) and related diapiric intrusion to zones of volume gain and passively emplaced plutons to zones of volume loss.

Pluton emplacement can also occur in extensional and compressional type shear zones. The best known example of emplacement in an extensional shear zone is the Strontian Granite (Hutton 1988a, b). Thrust type shear zones are associated with the Hercynian orogeny and Blumenfeld and Bouchez (1988) have described granites (mainly migmatite complexes and crustally derived leucogranites) emplaced in this regime. The Himalayan granites are another example (Le Fort et al. 1987) although the relationship between granite emplacement and the deformation has not been fully constrained.

The recent development of kinematic analysis of fault and shear zones has led to a greater understanding of the processes involved during crustal deformation. Hutton (1988a) envisages that the interaction between magma bouyancy and regional tectonics is necessarily complicated due to the anisotropies and viscosity contrasts which occur in the continental crust. "...crustal shear systems operate on an array of scales and in a complex linked network of failure planes and zones within which gaps and areas of dilation
may open transiently or permanently". He produced a general model in which
the emplacement style depends on the ratio of the tectonic extensional strain
rate (cavity opening) to the magma buoyancy rate. Passive mechanisms have
an excessive extensional strain rate and the magma is effectively sucked into
the opening void. Forceful mechanisms have an low extensional strain rate
relative to magma bouyancy rate.

1.5 The Ox Mountains Igneous Complex

The previous sections have briefly introduced the current theories on gran­
ite rheology, structural development and emplacement models. Much is known
although a lot is poorly constrained and conjectural. This thesis attempts to
provide additional information in an effort to solve some of the fundamental
questions which are not fully understood. Most of the emplacement mod­
els outlined above (section 1.5) discuss pluton emplacement on a large scale,
however this study has the advantage of access to several extremely well ex­
posed sections in which granitoid emplacement and deformation structures are
present. Detailed observation of these outcrops is combined with structural
mapping of the remainder of the pluton to create an emplacement model which
describes how this pluton is built up by the successive intrusion of its igneous
components. Rheological information from these exposures also provides de­
tailed examples of how granitoids behave during sub-solidus deformation.

It is necessary to provide a summary of the geology of the Ox Moun­
tains before outlining the questions which this thesis hopes to address. The
Ox Mountains Igneous Complex (OMIC) is emplaced entirely within the Ox
Mountains Inlier, Co.s Mayo and Sligo. The SW and central part of this in­
lier contains a series of lower amphibolite grade metasediments and volcanics
which have been correlated with the Middle Dalradian Rocks of Central Done­
gal and N.W. Mayo (Long & Max 1977, C.S. Jones Ph.D thesis, Durham 1989). Deformation in this inlier has been interpreted as part of the early history of the Highland Boundary Fault system (Hutton & Dewey 1986). The tectonic framework of the Ox mountains is consistent with the model of a major transpressive sinistral strike-slip shear zone (Chap. 2 for details). The syn-kinematic relationship between the transcurrent movements and the plutonic events have been established by Long & Max (1977), Jones (1989) and by this study.

This tectonic and magmatic framework provide access to several important questions concerning shear zone processes and granitoid rheology.

(1) What information does the granite deformation provide towards constraining the model of the evolution and geometry of the Ox Mountains mid-crustal shear zone and the early history of the Highland Boundary Fault?

(2) Can a deformational chronology be established throughout the pluton and can this be used to constrain the changing rheological response of the granitic rocks during sub-solidus deformation?

(3) What information do the microstructural features provide to suggest changes in the deformation mechanisms and how this might control the rheological response of the granitic rocks in the Ox Mountains?

(4) What do the structures, fabrics and internal contacts contained in the Ox Mountains convey about the mechanism for emplacement in the shear zone?

An outline of the tectono-metamorphic history of the metasediments which provide the wall rocks to the OMIC by Jones (1989) is summarized in the next chapter followed by a description of the OMIC components and the structures that have developed. A model for the emplacement is constructed and a discussion of the major conclusions which this study has produced and their implications for the above questions forms the remainder of this thesis.
CHAPTER 2

THE OX MOUNTAINS INLIER, SHEAR ZONE GEOMETRY AND GRANITE INTRUSION STRUCTURES.

2.1 Introduction

The Ox Mountains, also known as the Slieve Gamph Mountains, form an elongate, almost continuous (75Km) ridge from Castlebar (County Mayo) to Manorhamilton (County Sligo). This ridge rises above 350m along most of its length, reaching a maximum height of 544m at Knockalongy (G504 275).

The Ox Mountains topography reflects the fact that the rocks exposed comprise an inlier of 'old' pre-Caledonian and Caledonian metamorphic and igneous rocks which are more resistant to erosion than the adjacent Devonian and Carboniferous sedimentary rocks. The latter unconformably overlie the Ox Mountains Inlier to the SE and are in faulted contact to the NW (Long & Max 1977).

The Ox Mountains Inlier is divided into two blocks with differing lithologies and metamorphic histories (Fig. 2.1). The NE Ox Mountains comprise granulite facies psammites and basic rocks which are in tectonic contact with the central and SW Ox Mountains. This latter area contains a sequence of amphibolite facies, probably Dalradian age, metasediments and volcanics intruded by three plutons; the Ox Mountains Granodiorite (OMG), the Lough Talt Adamellite (LTA) and the Easkey Lough Adamellite (ELA).
Location and regional geology map

Fig. 2.1

Scotland
Ireland
Donegal Batholith
NE Ox
Ox Mountains Inlier
NW Mayo
L. Derg

0 50Km

Younger rocks
Caledonian Rocks
Pre-Caledonian basement
Caledonian Granites
2.2 Previous research in the Inlier

Previous studies in the Ox Mountains have been centred mainly on a debate concerning the age of the main tectonothermal event which the central and SW Ox Mountains experienced. This discussion has had implications for the age of the metasediments which are deformed by this event.

The earliest work was that by Symes (1872), who noted that a major fault, the Ladies Brae Fault, separated the SW and central Ox Mountains from the higher grade rocks in the NE Ox Mountains. Giekie (1893) concluded that this contact was the unconformable base of the Dalradian overlying Lewisian, whilst Kinahan (1878) and McHenry (1903) regarded the whole as being Cambro-Silurian.

The NE Ox Mountains high grade rocks have consistently been regarded as pre-Caledonian by all recent workers, however both pre-Caledonian, Moine equivalent and Dalradian have been suggested for those of the SW and Central Ox Mountains. A summary of the four recent interpretations is presented below.

(1) Phillips et al. (1975) suggested that the SW and central Ox Mountains rocks are a product of a pre-Caledonian retrogression of the high pressure granulites to the NE. They also considered that the entire Ox Mountains Igneous complex (OMIC) was intruded following the pre-Caledonian D2 event and that the intense flattening fabrics within the OMIC resulted from the final and third pre-Caledonian event. This was followed by D4 and D5 deformations which only deformed the low grade Cloonygowan Formation rocks (see Map 1). This latter unit was correlated into the Upper Dalradian.

(2) The rocks of the central and SW Ox Mountains have been correlated with the Appin and Argyll group Dalradian by Harris and Pitcher (1975) and Long (1974). Long and Max (1977) considered that the variety of lithologies,
particularly the presence of marbles and metavolcanics, were characteristic of the Dalradian and defined a coherent tectono-sedimentary succession. Long and Max (1977) did agree broadly with the deformation chronology erected by Phillips et al. (1975). This, together with their stratigraphic conclusions led to a model involving an early Grampian deformation event which was reworked, but only in the Cloonygowan Formation.

(3) Andrews et al. (1978) studied part of the central Ox Mountains and considered that the amphibolites, which Long and Max (1977) regarded as extrusive, were intrusive and emplaced at a late stage of D₂, prior to the mid-amphibolite facies peak (MP₂) of regional metamorphism. They regard the entire OMIC to be of Caledonian age and emplaced coevally with the D₅ and D₆ deformations during regional greenschist facies conditions. Therefore for them the Caledonian-Grampian cycle is younger than the D₁-D₄ pre-Caledonian cycle in the Ox Mountains sequence.

(4) Jones (1989) correlates the metasediments of the Ox Mountains with the Dalradian of central Donegal (Fig 2.2). He matched the structural history of the Cloonygowan Formation with the Ox Mountains Sequence recognizing an S₁ cleavage which predated the first two deformation events of Phillips et al. (1975) and Long & Max (1977). This produces a simplified chronological sequence of deformation events throughout the central and SW Ox Mountains, which is of Caledonian age. This is in tectonic contact with the granulite facies rocks of the NE Ox Mountains which have undergone pre-Caledonian deformation and metamorphism followed by Caledonian reworking.

The sequence of deformation and metamorphic events in the central and SW Ox Mountains is summarized in the next section.
Fig. 2.2

SOUTHERN DONEGAL
Alsop (1987)

SOUTHERN HIGHLAND GP
(UPPER DALRADIAN)

Croaghgarrow Fm
Shanaghy Green Beds
Mullyfa Grits
Agharan Fm.
Killeter Qtz

ARGYLL GROUP
(MIDDLE DALRADIAN)

L Eske Psammite / Termon Pelite
Boultypatrick Grit
Croaghubbrid Pelite
Reelan Fm.
Gaugin Qtz
Tillite
Crovenanta Fm.

CENTRAL AND SOUTHWEST
OX MOUNTAINS
Jones (1989)

CENTRAL AND SOUTHWEST
OX MOUNTAINS
Jones (1989)

Cloonygowan Fm

U Lismoran Fm.
Callow Member
L Lismoran Fm.
Ummoon Fm.
Newantrim Member
Leckee Transition Member
Leckee Quartzitic Fm.
Tillite
Tillite
petite member
limestone member

Approx Scale (all thicknesses are post tectonic)
(after Jones 1989)
2.3 Deformation in the Ox Mountains Inlier

The purpose of this section is to describe the deformation chronology, large scale structural geometry and kinematics which have been established throughout the SW and central Ox Mountains. This is essentially a summary of the work of Jones (1989)

2.3.1 D₁ : The first deformation

The original geometry of S₁ is unknown; it is therefore not possible to determine the geometry and kinematic significance of D₁. S₁ is a fine grained shape and alignment fabric, which is usually parallel to bedding but in some instances can be observed to cross cut it. It is often preserved in the hinges of the F₂ folds (Plate 2.1). The fine grained nature and composition of the S₁ fabric, which is expressed as a rarely preserved inclusion trail of muscovite and quartz in albite porphyroblasts, suggests that the metamorphic grade was in the low greenschist facies prior to the D₂ event. S₁ is likely to have been a weak fabric formed sub-parallel to the lithological banding, with no major D₁ structures formed in the Ox Mountains Inlier.

2.3.2 D₂ : The second deformation

Over most of the inlier the major D₃ event has transposed S₂ concealing the original geometry and kinematics of the D₂ event. D₂ was clearly important as it produced a strong penetrative fabric and minor F₂ folds. Evidence at Ardvarney (G152 954) demonstrates that the contact between the Cloonygowan formation and the Ox Mountains Succession was active at this stage.

D₂ deformation has resulted in a penetrative fabric in all of the metasedimentary and metavolcanic units in the Ox Mountains. This fabric is always observed parallel to any lithological banding. Unfolding the D₃ deformation suggests that S₂ was originally shallowly inclined and that F₂ fold axis were
orientated NE-SW perhaps parallel to the present $S_2$ strike.

Small scale $F_2$ folds are occasionally preserved, however the absence of more widespread vergence information means that the existence of larger scale $F_2$ folds can only be inferred from stratigraphic data. The only major $D_2$ structure identified by these means is the Lough Anaffrin Antiform (Map 1).

The intensification of $S_2$ and the progressive development of shear bands towards the contact of the Cloonygowan formation and the Ox Mountains Succession in the Ardvarney region strongly suggests that this is a tectonic contact of $D_2$ age. The Ardvarney area does not display the widespread development of the $D_3$ deformation and is likely to have been at some distance from the main locus of $D_3$ deformation prior to their tectonic juxtaposition. This may have been caused by the presence of a low angle detachment between the two areas or considerable vertical movement on the E-W fault which presently separates them. The same contact at Cloonygowan is presently expressed as a steep NE-SW trending $D_3$ shear zone which may have removed the original contact both horizontally and vertically.

There is no evidence for $D_2$ movement on any of the other tectonic contacts in the Ox Mountains. These contacts are also rheological boundaries between various formations of the Ox Mountains and are major $D_3$ tectonic slides. Intense $D_3$ deformation has removed any evidence for localized high strain during $D_2$ deformation.

Reconstruction of the $D_2$ geometry by Jones (1989) suggests an overthrust movement on the $D_2$ tectonic contact was towards a 275° azimuth.

2.3.3 $D_3$ : The third deformation

The present geometry of the Ox Mountains largely results from progressive $D_3$ sinistral, transpressive deformation which produces a kilometric strike-slip shear zone, whose development is described below.

Minor $D_3$ fold vergence indicates the presence of a major $F_3$ fold whose
axis is now sub-parallel to the spine of the inlier. This fold can be mapped at Lough Talt (G398 148) but is obscured by the OMG to the SW and by drift to the NW (Map 1). Minor F₃ folds are upright, close to isoclinal, with their axial planes parallel to the strongly developed steep S₃ cleavage and axis parallel to the gently NE or SW plunging stretching lineation (Plate 2.1). Evidence is provided in section 2.6.1 that the OMG is intruded syn-kinematically with respect to D₃ and can be used as a time marker to separate pre-intrusion geometries from those developed later in this progressive event. Minor F₃ folds, which are present in metasedimentary rafts, are more open than those in the country rocks, although their axes are still parallel to the stretching lineation.

The S₃ cleavage strikes NE-SW and dips steeply NW, or is vertical in most parts of the inlier (Plate 2.2). It is a spaced crenulation cleavage, which in higher strain zones transposes S₁ and S₂ into parallelism. It is intensified in localized high strain zones in which mylonite development is common. S₃ is parallel to the main foliation (S₁) developed in the OMG.

The strong mineral elongation or stretching lineation associated with this S₃ cleavage plunges gently NE or SW indicating a predominantly transcurrent component to the D₃ deformation.

In higher strain zones S₃ is deformed by a shear band cleavage or extensional crenulation cleavage, which indicates sinistral transcurrent movement (Plate 2.3). The development of this cleavage is associated with the widespread formation of S-C fabrics in the OMG (see 4.7) The development of sinistral shear bands in both the metasediments and the OMG is thought to take place at a late stage in the sinistral D₃ deformation.

The S₃ cleavage is locally intensified to form a braided system of high strain zones up to 200m wide which transect the Ox Mountains Inlier (Fig. 2.3). These high strain zones have been described as 'tectonic slides' by pre-
Plate 2.1.

Upright, close F3 fold, in Ummoon Formation semi-pelite with psammite bands refolding F2. \( S_1 \) preserved in the hinge of the latter fold. Burren Hill (G268 005).

Plate 2.2.

Large exposure of psammite and feldspathic semi-pelite of the Ummoon Formation which displays typical shear zone structures. The intense steep \( S_3 \) cleavage is axial planar to tight to isoclinal F3 folds. The view is approximately down the transport direction. (Cappagh M167 936).
Fig. 2.3

$D_3$ Sinistral shear zone

Tawnaneilleen shear zone

Oblique drop zone (Lough Easkey slide)

Frontal ramp (N Ox Slide)

Callow shear zone

Glennawoo shear zone

Approximate horizontal scale

0  10  20km

(after Jones 1989)
vious workers because they form along rheological boundaries which separate different formations of the Ox Mountains sequence. In the SW part of the inlier there are three such zones to the S of the OMG and one to the N which exhibit flat or straight sections, i.e. parallel to lithological contacts. These are the Glennawoo shear zone, the Lough Talt shear zone, the Callow shear zone and the Tawnaeilleen shear zone. Along strike to the NE, these shear zones diverge in the central Ox Mountains. The Glennawoo and Callow shear zones bend from a NE-SW orientation to E-W. A splay of the Glennawoo shear zone may bend to a N-S orientation passing underneath Easkey Lough. Here it is oblique to stratigraphy and separates the steep zone to the west of the Lough from a shallow zone to the east. These diverging shear zones are likely to have formed by a buttressing effect caused by movement on the tectonic contact between the NE granulite basement and the Dalradian rocks. This contact is a gently SW inclined overthrust to the N in which the Dalradian is carried over the basement. This movement is consistent with sinistral transcurrent movement on all the ductile high strain zones.

Towards the end of this sinistral shearing event the Knockaskibbole Fault system (see Map 2) becomes active and forms a series of dextral faults which overprint the sinistral fabrics. This produces a region in which greenschist facies dextral S-C fabric are dominant in the OMG at Dereens (G185 026). This structure is analagous in timing and kinematics with a break-back thrust.

2.3.4 D4: The fourth deformation

D4 structure reflect a fundamental change in kinematics that followed D3 and the intrusion of the OMG when transcurrent deformation ceased and uplift began throughout the inlier. D4 is everywhere represented by folds that have an angular or chevron geometry. The axes of F4 folds lie sub-parallel to the D3 stretching lineation and axial planes are moderately inclined in conjugate sets (Plate 2.4). There is no evidence of the development of an axial planar cleavage
Plate 2.3.

Sinistral extensional crenulation cleavages deform a mylonitic $S_3$ fabric developed in psammites of the Upper Lismoran Formation. (Callow, G324 026).

Plate 2.4.

$F_4$ folds deform the main $S_3$ fabric. The fold axis are parallel to the $D_3$ stretching lineation. The folds reflect a vertical shortening which followed the cessation of the main shear zone deformation. Easkey Lough (449 203).
to these folds. The F₄ folds have formed in response to vertical shortening and are best developed SE of Lough Easkey (G449 209). In this area the F₄ amplitude is up to 100m whereas it is usually less than 10m in the remainder of the Inlier. The Lough Easkey area is known to have undergone strong metamorphic retrogression which may also reflect rapid upward movement of this area. D₄ structures may reflect the uplift period which is known to have occurred between the intrusion of the OMG at c478ma and the adamellite plutons at c400ma (see chapter 5).

2.3.5  D₅ : The fifth deformation

The LTA and ELA plutons are situated adjacent to the trace of the Glen­nawoo shear zone, or a splay of this structure. Both plutons show evidence for solid state, sinistral deformation which may be the product of a c400ma reac­tivation of the Glennawoo shear zone. The plutons are emplaced post-D₄ and this reactivation is taken as being the D₅ event. This event was relatively weak compared with the earlier sinistral deformation, as there is no evidence for a late reworking of D₃ and D₄ structures in the Ox Mountains metasediments.

2.3.6  Terrane movements along the Highland Boundary Fault

The Ox Mountains Inlier is situated along the Fair Head - Clew Bay line which has been interpreted as the Irish continuation of the Highland Boundary Fault (HBF) (Max et al. 1983) (Fig 2.1). Deformation in the inlier has been interpreted as being part of early sidewall movements on this fault (Hutton & Dewey 1986). The HBF is considered by Soper and Hutton (1984) and Hutton (1987, 1989) to be a major terrane boundary in the British and Irish Caledonides. This structure has a protracted history; sinistral in early to mid-Ordovician, sinistral between end Silurian and early Devonian and dextral, reverse movements in post-late Devonian times. The main sinistral D₃ deformation described above is likely to be the mid-crustal expression of what was likely to be a Cordilleran style, allochthonous terrane boundary from
the Ordovician to the Devonian (Hutton & Dewey 1986).

2.4 The metamorphic history of the Ox Mountains

This section on the metamorphic history of the Ox Mountains Inlier is taken principally from the work of Andrews et al. (1978), Yardley et al. (1979) and Jones (1989).

2.4.1 The metamorphic assemblages

The metamorphic history is divided into three types by Yardley et al. (1979)(Fig. 2.4)

(1) High pressure granulite facies metamorphism in the NE Ox Mountains.
(2) Progressive greenschist to amphibolite facies metamorphism in the SW and central Ox Mountains.
(3) Retrogression of both (1) and (2).

Metapelites in the NE Ox Mountains contain:-
Kyanite + Kspar + garnet + biotite + plagioclase + biotite.
Metabasites are garnet pyribolites with pyroxene which is partly or totally altered to hornblende ± biotite. Phillips et al. (1975) estimate P-T conditions to be at least 10kbars and 800° C.

In the SW and central Ox Mountains the lowest grade rocks occur west of the Knockaskibbole fault, where metamorphism has not exceeded greenschist facies. Most pelites have garnet zone assemblages which include:-
garnet + biotite + chlorite,
garnet + chloritoid + chlorite,
paragonite + chloritoid + chlorite,
To the east of the Knockaskibbole Fault pelites develop:-

27
Fig. 2.4

Metamorphic rocks in the Ox Mountains after Yardley et al (1979)
garnet + staurolite + biotite + muscovite + plagioclase + quartz ± chlorite

East of Lough Talt the assemblage is:-

Kyanite + staurolite + garnet + biotite + muscovite + plagioclase + quartz ± chlorite

Yardley et al. (1979) cite evidence from a staurolite, pelite horizon with euhedral garnets which contain armoured, corroded relics of chloritoid parallel to quartz and opaque inclusions of an earlier foliation. This demonstrates the prograde nature of the metamorphism rather than its being a product of retrogression of NE basement granulites as suggested by Phillips et al. (1975).

Yardley et al. (1979) determined from 5 indicators that the peak of regional metamorphism occurred at approximately 600-620° C and 6-7kbars pressure with the temperature decreasing to 540° C towards the Knockaskibbole Fault.

The retrograde metamorphism occurs in the central Ox Mountains adjacent to the NE basement (Fig 2.4). Garnet and biotite are heavily chloritized and staurolite and kyanite are replaced by shimmer aggregates of fine muscovite, paragonite and margarite. Yardley et al. (1979) tentatively suggest that this region was also at amphibolite grade and experienced the same retrogressive metamorphism which also affected the NE granulites.

### 2.4.2 Timing of the metamorphism

Andrews et al. (1978) observed broadly similar assemblages in the central Ox Mountains to those described by Yardley et al. (1979) and considered the peak of regional metamorphism to be MP₂, since kyanite porphyroblasts are augened by the main fabric. An earlier S₁ fabric is preserved as crenulated inclusion trails of muscovite, quartz, graphite and opaques in MP₂ porphyroblasts. Yardley et al. (1979) stated that the metamorphic peak was formed after the main pervasive S₂ fabrics which is is in agreement with the work
of Jones (1989). Jones describes similar assemblages to those of Yardley et al. and Andrews et al. and notes that kyanite porphyroblasts from Zion Hill (G424 174) are deformed by the main D₃ fabric.

2.4.3 The thermal aureole of the OMIC

The OMG was considered to have been emplaced syn- to late D₂ by Max et al. (1976) and Long & Max (1977). Their D₂ can be directly correlated with the D₃ described by Jones (1989) as outlined above. The OMG is not thought to have an appreciable thermal aureole (Long & Max 1977, Yardley & Long 1981). The latter authors suggest three possible reasons for this.

1. The lithologies adjacent to the contact are mainly psammite and semi-pelite and are unresponsive to further heating.

2. The wall rocks were still hot after the kyanite grade peak of regional metamorphism.

3. The OMG was intruded by a 'relatively cool tectonic emplacement'.

It is unclear to the present author precisely what a cool emplacement mechanism is unless Yardley and Long (1981) refer to the emplacement of an entirely crystalline body of granodiorite. Ample evidence of thin dykes and the internal sheeted nature of the granodiorite suggest that the OMG was mobile and hot during its emplacement. Taylor (1966) reported that the highest temperature regional metamorphic zones where found adjacent to the OMG which would suggest that the granodiorite was emplaced close to the regional metamorphic peak. During the course of this study and that of Jones (1989) abundant sillimanite was discovered at one locality, in pelitic rocks which are situated in a sheeted granodiorite and country rock contact zone (Plate 2.?). This locality is 200m NW of the N end of Callow Lough lower (G306 043) and approximately 300m SE from the main body of the pluton. Its discovery provides evidence that the granodiorite was able to provide substantial heat to the wall rocks but that this effect was limited to a narrow zone around the
pluton.

The LTA and ELA produced a pronounced thermal aureole in the country rocks adjacent to their contacts and this is described in relation to each pluton in chapter 5.

2.5 Intrusion related structures in the OMG

2.5.1 Timing of emplacement relative to the deformation chronology

The timing of the intrusion of the OMG has been established by this author using two main methods:—

(1) Examination of the deformation of minor granodiorite intrusions in the sheeted contact zones, particularly at the NW contact and at Callow, Attimachugh and Ummoon on the S of the pluton.

(2) Observation of the deformation preserved in rafts of metasediments which become enveloped in the granitic rocks during intrusion.

(1) Structures in the sheeted wall rock zones.

At Boyhollagh (G307 101) there is evidence for the disruption of the semipelitic and psammitic country rocks by brittle faulting with fault planes dipping more gently to the NW than the main S₃ fabric in the metasediments. This faulting produces a series of cutoffs of S₃ which have a sinistral transcurrent component and also an overthrust component to the SE. Some of these thrust faults contain granodiorite veins (Fig. 2.5) and they offset the S₃ cleavage which is axial planar to tight F₃ folds. In other situations, these structures are modified by further ductile sinistral shearing associated with D₃, producing a series of boudins, and then folded by F₄ to form conjugate folds with chevron type geometries (Fig. 2.5). This implies that intrusion took place during the D₃ event.
Figure 2.5.

Minor intrusion related structures developed in the contact zone on the northern flank. Scale bar =1m. Boyhollagh (G307 101).
Fig 2.5

(a) Upright \( F_3 \) truncated by brittle oblique overthrust

(b) Open folding of Granodiorite

(c) \( D_3 \) pervasive shearing continues forming boudins

- Axial planar cleavage parallel to \( S_3 \)
- Granodiorite veins folded by \( F_4 \)
In the sheeted zone, 200m NW of the shore of Callow Lough lower (G306 043) there are several sheets of megacrystic granodiorite, one is up to 100m thick whereas most are less than 10m. These granodiorite sheets contain diorite xenoliths and pods of appinitic material typical of the granodiorite facies along the S margin of the pluton (see chapter 3). These sheets are intruded sub-parallel to the 045-052° strike of D3 fabric in the semi-pelites and contain a co-planar S1 granodiorite foliation. Most outcrops in this area are flat surfaces, however at Hill 376 (G305 039) a small 3m cliff, perpendicular to strike permits a view of the granodiorite sheets in three dimensions (Plate 2.5). In this cliff, it is apparent that the granodiorite contact is 15° oblique to the S3 foliation which is parallel to S1 in the granodiorite. These field relations suggest that the granodiorite was emplaced prior to the end of D3 deformation but it is not possible to say if it occurred during or before D3 from this area. There are numerous small scale high angle reverse shears which indicate overthrusting of both granodiorite and metasediments to the NW.

Sheets of granodiorite are very common along this SE margin and are particularly well developed at Attimachugh (G323 062), Curranara (G283 023) and at Crumlin (M170 972). In these zones, intrusion has occurred prior to the end of the D3 event as the sheets all contain the S1 foliation parallel to the main shear zone cleavage.

A small intrusion of granodiorite occurs 200m S of Burren Hill (G268 005) and although this body is situated 1.3km S of the main part of the OMG, it is identical in terms of composition, texture and its microstructural characteristics. This intrusion is elongate, 570m by 80m, in dimension and length parallel to the main S3 foliation. There are a series of brittle minor structures adjacent to the contact with the metasediments which display brittle overthrust type characteristics and granodiorite intruded along active thrust planes (Fig. 2.6).
Figure 2.6.

Field sketch of contact relationships between Group 2 granodiorite and Leckee Quartzite. Burren Hill (G207 004).
many cutoffs of the $S_3$ fabric in the overturned limb of fold pair

out of footwall syncline thrust

Leckee Quartzite Formation

granodiorite dykes emplaced along active thrust planes
These granodiorite sheets subsequently acquire an \( S_1 \) foliation which is parallel to the \( D_3 \) fabric in the metasediments and also show evidence for open, upright folding in a similar style to \( F_3 \) folds. The evidence from this area suggests that granodiorite was emplaced in a brittle manner during progressive \( D_3 \) deformation. Significantly, this small intrusion is situated along the trace of the Lough Talt shear zone, or tectonic slide, which separates the Lec-kee Quartzite from Ummoon formation semi-pelites (see chapter 7 for further discussion).

(2) Structures contained in the metasedimentary rafts.

All metasedimentary rafts in the Ox Mountains contain a strong, steep fabric which is similar to the \( S_3 \) developed in the wall rocks. In many cases, particularly in psammitic rafts, the fabric is strongly recrystallized and the rafts have a hornfelsed appearance. Many rafts contain minor folds which are identical in structural style and orientation to \( F_3 \) folds in the wall rocks. In the rafts fold axes are parallel to the gently plunging stretching lineation and the folds have axial planar \( S_3 \) fabrics. This is parallel to \( S_1 \) in the surrounding granitic rocks. One example from the shore section at SW Lough Cullin (G223 011) displays these features particularly well (Plate 2.6). Here an upright, probably \( F_3 \) antiform, with its axis parallel to the stretching lineation is enveloped by granodiorite and granite. The igneous contact which overlies the raft is not folded indicating that the \( F_3 \) formed prior to the intrusion. However, the \( S_1 \) foliation and stretching lineation in both igneous units are co-planar and collinear respectively with the \( D_3 \) structures in the raft suggesting that \( D_3 \) continued after the intrusion had occurred.

There are other less clear examples in the OMG particularly those developed along the NW flank between Carrowdoogan (G307 129) and Bunny-connellan East (G365 164). The steep \( S_3 \) fabric in all rafts is axial planar to upright \( F_3 \) folds, which are in general more open than the almost isoclinal \( F_3 \)
Plate 2.5.

Sheets of Group 2 granodiorite are intruded into the semi-pelites and psammites of the Leckee Quartzite Transition member. The sheet contacts are 15° oblique to the main S3 fabric in the metasediments with the intersection being sub-horizontal. The granodiorite S1 foliation and stretching lineation are coplanar and collinear with the respective D3 structures. Hill 376 (G305 039).

Plate 2.6.

Upright, tight F3 fold in a metasedimentary raft is enveloped by granodiorite and biotite granite. Contact between biotite granite and granodiorite (above map in photograph) is not folded implying that the F3 was formed prior to magmatic intrusion. The foliation and stretching lineation in the granitic rocks are exactly parallel to the D3 structures indicating that the shear zone deformation continued after the granite emplacement event occurred. SW Lough Cullin (G212 011).
minor folds in the wall rocks. This indicates that they did not undergo fold limb tightening in the final stages of D3. This deformation is likely to have been concentrated in the formation of the planar S1 and S-C fabrics in the granitic rocks.

The syn- to late D3 age indicated by the intrusion related structures agrees with the interpretation of Taylor (1969), Max et al. (1976), Pankhurst et al. (1976), Long & Max (1977) Long et al. (1984), Max & Long (1985) and Jones (1989) but disagrees with that of Phillips et al. (1975) and Andrews et al. (1978), who interpreted the OMG emplacement age as post-D4. Their age relationships may have come from the LTA and ELA plutons which are post-D4 intrusions (see chapter 5) and were originally thought to be the same age as the OMG. The OMG has been dated using the Rb/Sr whole rock method. This produces an age date of 478 ± 12 Ma (Pankhurst 1976) with an initial \(^{87}\text{Sr}/^{86}\text{Sr}\) ratio of 0.7056. This age date can be used to constrain the early history of the Highland Boundary Fault. Sinistral movements on this fault system are thought to have occurred during the Ordovician, and this is supported by the Ox mountains Granodiorite age date and field relations.

2.5.2 The nature of the OMG intrusion related structures

The timing relations between OMG intrusion and the regional deformation chronology, outlined above, indicate syn-D3 emplacement. The D3 event is characterized by intense sinistral shearing which takes place in a ductile manner. In areas away from the granodiorite, there is little evidence for brittle deformation and the S3 fabric is folded by conjugate D4 folds which indicate vertical shortening. However in localities close to the granodiorite contacts there is evidence for a change in deformation style from ductile pervasive shearing to a more gently inclined brittle overthrusting. This is combined with granodiorite intrusion in a series of slightly oblique sheets or dykes which have sharp contacts, indicating a brittle intrusion style. These sheets
are subsequently overprinted by a continuation of the pervasive ductile sinistral shearing which forms the $S_1$ foliation, S-C fabrics and asymmetric boudins in the contact zones. These structures are folded by the conjugate $F_4$ folds. The observation of these structures indicates a fundamental change in deformation style during D$_3$ which is likely to be related to the presence of granitic melt in the shear zone during D$_3$ (see Chapter 7).

The granodiorite intrusion at Burren Hill (G207 004) illustrates the change in deformational style (Fig. 2.6). Here a NW verging fold pair display numerous high angle thrusts which displace part of the lower and middle limb of the fold pair up to the NW. Granodiorite veins are also intruded along some of the thrust planes and are folded by the thrusting. Figure 2.7 is an interpretation of this structure in which a larger granodiorite sheet below has intruded along the axial plane of the fold pair and has pushed fingers of granodiorite along thrust planes ahead of the main body. It may be postulated that the magma and its associated fluids induced brittle thrusting within amphibolite facies ductile shearing because of the localized high fluid pressures associated with the magma. The high intrusion stress rate could not be accommodated by ductile folding causing a change to brittle mechanisms. The importance of the brittle emplacement mechanism relative to the emplacement model for the OMG is discussed more fully in Chapter 7.
Ductile strain rate not fast enough to accommodate ascending granodiorite strain rate
This changes deformation to a brittle mechanism

Interpretation of Fig. 2.6
CHAPTER 3

THE OX MOUNTAINS GRANODIORITE COMPONENTS

3.1 Introduction

The OMG comprises four major components: granodiorite, tonalite, muscovite granite and diorite. In the following sections the contact relations and petrography of each of these units is described, compared and contrasted. There are also four minor components which are briefly described. Each division of the major components include a range of compositional and textural variations which were established during field investigation and has been confirmed during subsequent microscopic investigation of representative samples.

Table 3.1 provides results from the modal analysis of various samples of the four major components of the OMG. The data have been calculated from point counting of 47 thin sections which were considered to represent typical samples of the four units. The plagioclase, alkali feldspar and quartz components have been recalculated to 100% and plotted on a Streckeisen (1976) diagram which is a method of classifying granitic rocks (Fig. 3.1). The Ox Mountains samples show a range of mineralogical variation which is wider than that in most plutons and this raises the question as to the origin and evolution of the magmas. It is not an object of this thesis to attempt to explain the petrogenesis of the OMG, however detailed sampling of the components has taken place to provide major and trace element geochemical data which will attempt to constrain the origin and evolution of the magmas. This work was undertaken in conjunction with R.J. Reavy (University of Oxford) and results will be available at some future date. Possible origins for the magmas
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<th>DIORITE</th>
<th>BIOTITE GRANITE</th>
<th>MUSCOVITE GRANITE</th>
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</tbody>
</table>

Table 3.1
Fig. 3.1

AFTER STRECKEISEN (1976)

47 samples

quartz

alkali feldspar granite

syenogranite

monzogranite

granodiorite

alkali feldspar syenite

quartz syenite

quartz monzonite

quartz monzodiorite

diorite

alkali feldspar syenite

syenite

monzonite

monzodiorite

diorite

7 samples of diorite

plagioclase
will be discussed in Chapter 7 in which the emplacement model for the OMG is discussed.

3.2 The Ox Mountains Granodiorite

The Ox mountains granodiorite is volumetrically the most important component exposed in the pluton. It can be mapped continuously from Crumlin (M164 969) to Bunnyconnellan East (G369 166) (Map 2) and is considered to be approximately 100Km² in area.

3.2.1 Contact relations of the OMG

The granodiorite contact zone is best exposed along the SE flank in a 4km belt from Leckee (G267 022) to Callow Lough Lower (G306 042). Here the contact is transitional and sheeted with a change from granodiorite sheets in the Leckee Quartzite to meta-sedimentary rafts in the pluton. Granitic veins and sheets are observed up to 1.5Km outside the contact and metasedimentary rafts are common up to 750m inside the pluton. A typical exposure of megacrystic granodiorite sheeted in metasedimentary rocks of the Leckee Quartzite formation is exposed 350m SW of the northern end of Callow Lough Lower (see Plate 2.5). Here 12 granodiorite sheets up to 1m thick intrude the semi-pelites and psammites in a zone which is about 200m wide.

The contact zone is not well exposed in the region from Muckanagh (G233 002) to Crumlin (M179 975) although psammites and quartzites with granodiorite veins on one side are separated from granodiorite outcrops within the pluton by an exposure gap up to 750m wide. At Derryhick Lough (M205 990) this exposure gap is reduced to 400m. A sheeted zone is present along strike to the NE at Attimachugh (G323 058) with up to 20m wide rafts of metasediments (mainly quartzites) and sheets of megacrystic granodiorite. Further N there is an area of no exposure from Knockfadda (G342 078) to the contact
region of the poorly exposed Lough Talt Adamellite (LTA).

The North contact displays more complicated relationships in which granodiorite is intruded adjacent to outcrops of muscovite granite as well as Attymass Formation metasediments present on the northern flank of the inlier. At Stoneparkbrogan (G235 072) a fault separates the muscovite granite from granodiorite. A metasedimentary screen is also situated along this fault line and this obscures the age relationship between the granite and granodiorite in this region. No veins of granodiorite were recognised in the muscovite granite or (vice versa) but the relative strain intensity of the foliations in both units suggests that the much more strongly deformed muscovite granite is the oldest (see section 3.6.1).

At Bunnafinglas (G288 098), on the northern contact the muscovite granite is absent and the granodiorite displays a concordant contact with semipelites and psammites of the Attymass formation. A planar foliation is strongly developed in the granodiorite rocks within 20m of the contact. The granodiorite contains a little muscovite within a 5m zone contact zone, but does not display the strong white weathering typical of the muscovite granite which may be due to a greisening.

Further N at Derryvicneill (G291 097) there is a 250m exposure gap between a tonalite body, which is in contact with and separates the main granodiorite from the metasediments which are sheeted with muscovite granite. 500m to the N at Boyhollagh (G314 106) a similar relationship is displayed.

At Carrowdoogan (G315 119) good exposure is provided in a 1Km stream section through the the N flank of the pluton. The actual contact lies in a 400m exposure gap which can be found at the prominent break in slope between the granodiorite and the metasediments in this region. Figure 3.2 shows a detailed log through this stream section at Carrowdoogan and this illustrates some important features of the NW contact zone. A large granodiorite outcrop just
Figure 3.2.

Summary geological map of a 1Km stream section through the northern flank of the pluton. Carrowdoogan (G315 119).
Fig. 3.2

N

0 10 20 30m

wall

no outcrop

Fig. 3.2

granodiorite
biotite granite
Ummoon Fm. semi-pelites
Leckee Transition Fm. psammites

massive granodiorite, quite biotite rich with large elongate quartz ribbons

granodiorite boulders

biotite rich granodiorite strong foliation

and sinistral SC fabric

biotite granite

Leckee Transition Fm. psammites

biotite rich psammites with upright folds

uniform granodiorite

Muskovite granite sheet

biotite granite sheet

biotite rich granite with quartz segregations and 1m muscovite granite sheets

semi-pelite raft contains psammites and quartz segregations and 1m muscovite granite sheets

mainly semi-pelite with quartz segregations and 2-3cm psammite bands with 2mm albite porphyroblasts

mainly semi-pelite with quartz segregations and 2-3cm psammite bands with 2mm albite porphyroblasts

biotite granite sheet

granodiorite with alkalis feldspar phenocrysts

dark semi-pelite raft thin psammite bands

overhanging cliff of granodiorite

3m cliff of granodiorite with semi-pelites large quartz ribbons

semi-pelite raft with quartz segregations folded by upright folds

2m thick semi-pelite with quartzite

1m semi-pelite with quartzite

semi-pelite rafts with thin biotite rich psammites
above the break in slope can be mapped in an upstream direction for 150m on a boulder-strewn slope. A series of elongate rafts of metasediments are exposed in the next 250m. These rafts range in widths from 5-15m and their lengths were not determinable (70m minimum) and they are composed mainly of semi-pelites with thin (2-3cm) psammite bands which contain albite porphyroblasts. The contacts between the rafts and granite are broadly concordant and the granodiorite and metasediments foliations and lineations are coplanar and collinear respectively. The rafts contain early quartz segregations and upright folds parallel to the stretching lineation which are considered to be typical of the shear-zone deformation.

A 25m semi-pelite raft separates this granodiorite and metasedimentary raft zone from a biotite granite and metasedimentary raft zone. Moving upstream from this granite sheet the next raft encountered is a 100m wide semi-pelite and psammite raft which contains 2cm thick amphibolite sheets and several 1m thick muscovite granite sheets up to 1m thick (see 3.6.1 for discussion on the relative ages of these granites). Further upstream the exposure is not as good but isolated outcrops of biotite granite occur in the next 80m. Towards the top of the stream section is a 100m thick exposure of a biotite, epidote-rich psammite which again contains the upright folds and is considered to be a transitional unit between the Leckee Quartzite Formation and the Ummoon or Attymass Formation. The contact between this and the biotite granite is not exposed. At the top of the stream a large outcrop of granodiorite is present and this is separated from the psammites by a small fault along which the stream has cut a shallow gorge. The psammites adjacent to this fault appear slightly brecciated.

This stream section provides an illustration of several important features of the OMG.

(1) The internal sheeting of the pluton can be observed; in this case be-
between muscovite granite and granodiorite.

(2) These sheets disrupt what may be an original stratigraphic package of metasediments, i.e. the pluton has a ghost stratigraphy (Pitcher 1970, Pitcher & Berger 1972)

(3) It provides evidence towards establishing the age relationships of some of the major igneous components (see 3.6.2).

At Graffy (G322 124) the granodiorite can be mapped in concordant contact with metasediments which dips 56° to the NW. These metasediments contain concordant muscovite granite sheets 110m outside the main contact. At Ellagh Beg (G331 133) the contact with the metasediments is not exposed, however a possible continuation of the muscovite granite sheet at Graffy has been mapped here, again concordantly dipping moderately to the NW. In this townland there are several metasedimentary rafts which show varying degrees of discordance at their contacts. Moving in from the concordantly intruded muscovite granite sheets, there are several isolated examples in which the metasedimentary raft foliation dips much more gently than the granodiorite foliation and the raft contacts are parallel to the raft cleavages. 1.5km from the contact there is a large 300m wide raft at Winny Langans Lough (G343 134) which again shows concordant relations between the granodiorite contacts and the metasedimentary foliation with both dipping 62° NW. The vertical height distance between the contact and Winny Langans Lough is approximately 200m enabling the construction of a sketch section (Fig. 3.3). This shows a possible interpretation of the contact relations. This reconstruction implies a component of vertical uplift of the contact zone relative to the central parts of the pluton. The steep foliation in the granodiorite represents the results of a superimposed, heterogenous, deformation by the shear zone which left the gently inclined raft contacts in their present orientation.

At Bunnyconnellan East the contact is exposed between Attymass forma-
Fig. 3.3

Original $F_3$ fold structure intruded by granodiorite

main foliation in granodiorite

concordant relations at Winny Langans Lough

concordant contact at Ellagh Beg

rafts with gently inclined contacts and fabrics

horizontal scale = vertical scale

0 150 300m
tion metasediments and granodiorite however it appears modified by a late fault because a narrow breccia zone exists in the granodiorite at the contact. This results in the granodiorite and metasediments being discordantly disposed in this area, with the semi-pelites dipping gently to the NNW and the granodiorite moderately to steeply to the NW. There is a large straight valley along the extension of this contact which could a fault valley.

The granodiorite is in faulted contact with a major muscovite granite body at Knocknasliggaun (G371 150) but the nature of the contact is not obvious. In the breccia there are fragments of metasediments (mainly semi-pelites and calc-silicates) suggesting that there may have been a metasedimentary screen along this contact which was subsequently faulted.

The pluton is truncated at the NE end by a major fault trending 125° which passes through the Gap (G372 161) and Lough Talt. The movement sense on this fault is not known but is suspected to have a component of downthrow to the SW. The contact between the granodiorite and the Lough Talt adamellite is not exposed.

The SW contact of the OMG is also controlled by fault movements. The Knockaskibbole Fault (Long & Max 1977) which trends 029° truncates the pluton and brings metasediments into a discordant contact with the granodiorite. This fault also passes through the pluton and produces a large brecciated zone between Terrybaun (G191 953) and Largan (G179 021) (see section 4.10).

3.3 The petrography of the granodiorites

Hand specimen examination of the granodiorite indicates variations in colour index and texture. The granodiorites can be divided into two groups principally on their textural variations.
3.3.1 The Group 1 Granodiorites

Group 1 granodiorites (plate 3.1) are equi-granular, medium grained light to medium grey and comprise white plagioclase, pink alkali feldspar, greenish-brown biotite and colourless quartz. The average grain size is 4-5mm, and they form the major part of the northern and central regions of the pluton (See Map 2).

Plagioclase crystals are equant, subhedral and measure up to 6mm long. The composition varies from An$_{32}$ to An$_{20}$. Most of the plagioclase displays normal continuous zoning and the twinning is mainly albite and carlsbad, although two sets of albite twins at right angles occur in rare examples. Plagioclase inclusions are equant biotite, epidote, anhedral chlorite and rare small flakes of muscovite. In most cases the plagioclase displays alteration to sericite.

Alkali Feldspar is present as inequant, subhedral crystals up to 6mm long. Both orthoclase and microcline are present in the same specimen. Orthoclase is commonly microperthitic and untwinned although carlsbad twins do exist in some samples. Both orthoclase and microcline contain euhedral biotite and epidote inclusions towards their centres. Plagioclase inclusions do occur along the margins but these may be related to solid state deformation and reaction (see section 6.3).

Biotite is present as elongate (up to 1mm) subhedral laths which often form aggregates up to 5mm in length. These therefore contrast with the euhedral biotite inclusions in the feldspars. The biotites contain zircon inclusions which have pleochroic haloes and blebs of opaques. The biotite in many cases shows alteration to chlorite.

Quartz is in the form of elongate aggregates of grains which are up to 1cm long and 1mm wide. Individual grains show evidence of being highly deformed and show undulose extinction (see section 6.3).

Minor components include chlorite which is present as anhedral aggregates.
and may be a late alteration product. There are rare hornblendes (especially in the southern part of the pluton) in slightly more mafic varieties of the granodiorite. These are equant, subhedral up to 0.2mm and are associated with mafic aggregates of biotite, epidote and chlorite. Muscovite is present along some alkali feldspar grain boundaries and is associated with sericite in the plagioclase crystals.

Accessories include euhedral epidote (0.1mm) inclusions in plagioclase and larger (0.5mm) anhedral epidote which is present outside the phenocryts. Zircons are small (0.1mm) and present in biotite. Small allanites and apatites have been observed, as have very rare clinozoisite crystals. Opaques occur as blebs in biotite and at the margins of feldspar phenocrysts.

Closely associated with the Group 1 granodiorites is a series of granites (Plate 3.2) which display a similar equigranular textures in hand specimen as the granodiorites. The Group 1 granodiorites are intruded into the granites at Carrowdoogan (G322 117)(Plate 3.3). The granites show an increase in the alkali feldspar content accompanied by a decrease in plagioclase and mafic content. The alkali feldspar is mainly microcline and is microperthitic sometimes displaying Carlsbad twins. The other phases closely resemble those described for the group 1 granodiorites. These granites have been named the biotite granites to distinguish them from the muscovite bearing granites which occur along the N contact (section 3.6).

There are also some minor occurrences of a hornblende rich granodiorite in which large (2cm) aggregates of hornblende are formed in small diffuse patches especially on the SW shore of Lough Cullin (G212 011).

3.3.2 The Group 2 granodiorites

Group two granodiorites are porphyritic with pink alkali feldspar megacrysts up to 6cm long in a medium grey groundmass comprising white plagioclase, pink alkali feldspar, colourless quartz, brown biotite and acces-
Plate 3.1.

Hand specimen of the Group 1 granodiorite which occurs throughout most of the northern part of the pluton. The major phases are plagioclase, alkali feldspar, biotite and quartz. Derryvicneill (G279 092).

Plate 3.2.

Hand specimen of the Biotite granite which is associated with the Group 1 granodiorites. It has an equigranular texture and is principally composed of alkali feldspar, plagioclase, biotite and quartz. Pontoon Dancehall (G201 040)
sories. These megacrystic granodiorites have been mapped in an elongate region close to the S contact (Map 2) however some small areas of megacrystic granodiorite do occur along the N contact, e.g. Cliff Island (G215 054) although here the alkali feldspar phenocrysts here are rarely larger than 3cm (Plate 3.4).

The main distinguishing feature between the Group 2 and the Group 1 granodiorites is the presence of the alkali feldspar megacrysts which are characteristically pink in hand specimen. These are elongate, subhedral orthoclase crystals which are generally untwinned although Carlsbad twinning does occur. The orthoclase crystals are microperthitic in some cases and there are examples where microperthitic orthoclase has undergone a partial inversion to microcline. The orthoclase megacrysts contain euhedral epidote and subhedral biotite as well as chlorite inclusions up to 1mm in size. Plagioclase inclusions occur close to the margins of the orthoclases (see section 6.3) The orthoclase and microcline is not visibly altered.

Plagioclase is also believed to have been a phenocryst phase in group 2 granodiorites and is present as elongate, subhedral crystals up to 5mm long. These display albite and Carlsbad twinning although unzoned examples are not uncommon. Plagioclase composition varies from An29 to An23 and crystals display a normal, discontinuous zoning. Inclusions present are, euhedral epidotes and equant euhedral biotites which are smaller than those in the orthoclase megacrysts. The plagioclase is variably altered to sericite.

The groundmass is a medium grained granodiorite which in some cases could be classified as tonalite due to its very low alkali feldspar content. The subhedral, equant plagioclase has a composition which averages An23 and displays normal continuous zoning. It is altered to sericite and does not contain any inclusions. The alkali feldspar is mainly orthoclase which is microperthitic.

Biotite in the groundmass is present as greenish brown elongate aggregates
Plate 3.3.

Granodiorite has been intruded as narrow sheets into the biotite granite. The granodiorite has pushed two 'fingers' into the biotite granite, forcing it apart and creating a bridge of granite between the two dyke tips. Carrowdoogan (G315 119).

Plate 3.4.

Hand specimen of Group 2 granodiorite which displays alkali feldspar phenocrysts in a groundmass of biotite, plagioclase and quartz. This is typical of the granodiorite facies formed along the southern contact and in rare outcrops on the northern contact. Cliff Island (G215 054).
up to 3mm long composed of individual biotite laths which are 0.5mm long. The biotite shape is subhedral in contrast to that in the phenocrysts. It is altered to chlorite and opaques although the degree of alteration is variable in the sample set.

Quartz aggregates are elongate and up to 6mm long. They are composed of equant anhedral quartz crystals which are approximately 0.5mm in diameter. Most crystals show undulose extinction and have wavy grain boundaries indicating that they are sub-grains which have formed in response to an external stress field (see section 6.3).

Other minor phases which are present in the group 2 granodiorites include anhedral chlorite which may be a replacement product. Euhedral epidotes form inclusions in the phenocryst phases whereas in the groundmass they are anhedral. See section 3.9 for discussion of the inclusions in relation to the possible crystallization sequence for the granodiorites and section 3.5 for the importance of euhedral epidote inclusions.

3.4 The Ox Mountains Tonalites

The Ox Mountains Tonalites form linear dykes zones up to 8km long and 800m wide. These zones are elongate sub-parallel or slightly oblique to the long axis of the pluton. They are exposed predominantly in the SW of the pluton in the area from Lough Cullin to Rosses West (Map 2) although there are several examples to the east of Lough Cullin. The tonalites are intruded into the granodiorites and biotite granites described in the previous section. In the SW part of the pluton there are three main zones (see Map 2).

3.4.1 Zone 1 tonalites

In the south of the pluton a tonalite zone can be mapped discontinuously for 8 km from Crumlin village (M173 976) to the SW shore of Lough Cullin
at Bunduvowen (G221 019). It is exposed in three main elongate bodies along this zone. The zone width is variable and generally widens towards the NE. It is 250m wide at Crumlin, 500m thick at Cunnagher North (M190 988) and 800m thick at Bunduvowen.

The Crumlin body is 700m long and its boundaries are swarms of linear, interdigitated discrete dykes and granodiorite screens. At Crumlin the dykes dip 50-60° NW varying between 20-60cm in width and are intruded on both the N and S boundaries into the group 2 megacrystic granodiorite which here also contains dioritic material.

The Cunnagher North body is much more elongate and is 3km long. Exposure is poor, so it is not certain if this body is continuous or composed of two separate bodies which are intruded along strike from each other. The dyke swarm on the north boundary dips 54 - 63° NW and is composed of 2-3m wide tonalite dykes intruded into granodiorite over a width of 100m in which tonalite makes up approximately 30% of the rock area. The southern boundary contains narrower dykes (average 50cm) emplaced within a 50m belt of group 1 granodiorite which also contains psammite rafts.

A narrow tonalite body is exposed 500m S of the Bunduvowen body. It is 700m long by 100m wide and contains up to 3m wide granodiorite screens. Tonalite dykes are also present on the SW shore of Lough Cullin (G212 011) directly along strike from this body. This region provides an excellent section through this part of the southern region of the pluton. Tonalites are exposed in a 300m section which trends NNE-SSW (Fig. 3.4). Tonalite makes up 32% of the total area exposed in this section. The largest body is 90m wide and contains micro-diorite and gently inclined micro-granite sheets. Within this section the minor tonalite sheets display slight variations in their composition and textures. The other tonalites here vary from 1-10m in width and are intruded into Group 2 granodiorites, some diorite intrusions and psammitic
Figure 3.4.

Geological map of rocks exposed on the SW shore of Lough Cullin (G223 012-G224 009).
Fig. 3.4

- Deformed pegmatites
- Uniform tonalites with faint banding
- Flat lying microgranite sheets
- Upright folds of main foliation
- Amphibolite dyke
- Microgranite dyke with xenoliths
- Discrete dextral shear zones deform granodiorite/tonalite contacts
- Quartzites with upright folds axis parallel to stretching lineation
- Granodiorite
- Tonalite
- Microgranite
- Microdiorite
- Quartzite
- Pegmatite
- Amphibolite
and quartzitic metasedimentary rafts.

At Bunduvowen, the total width of the tonalite is 800m with dyke swarms on both the northern and southern boundaries. There is however, a septum of group 1 granodiorite in the centre of this tonalite which may divide it into 2 small adjacent bodies. The north boundary is a 100m wide dyke swarm dipping 56-70° NW in which 2-4m tonalite sheets comprise approximately 40% of the area. This boundary is gradational between granodiorite and tonalite. The south boundary contains a finer scale dyking over a 100m belt in which 10-50cm tonalite dykes and Group 1 granodiorite screens in equal proportions can be mapped in many areas. Uniform outcrops of Group 1 granodiorite separate these finely sheeted zones.

The Bunduvowen body is offset by 100m in a left stepping manner from the NE end of the Cunnagher N. body. The intervening granodiorite between the tip areas of these bodies contains tonalite dyking although the scale is reduced as is the volume of dyke material. The Cunnagher North body is offset on a left stepping manner along its length and the Crumlin body is also offset in a left stepping manner from the Cunnagher North body.

3.4.2 Zone 2 tonalites

Zone 2 tonalites occur in a broad zone from Greenauns (G170 004) to the W. shore of Lough Cullin (G208 028). This zone does not display the same continuous bodies of tonalite as in zone 1; instead it is mainly composed of isolated elongate bodies in which the maximum length is 2km, although most are less than 1km. The maximum width is 400m. Two bodies are present at Greenauns. The western body is exposed 200m S of Greenauns School (G182 009) and is an elongate discontinuous body which occurs as a prominent NE-SW ridge which is 400m long and 100m wide. It is a dyke/host rock sheeted body throughout in contrast with zone 1 tonalites which are homogeneous in their centres. The tonalite dykes here are up to 2m wide, approximately 20-
30m long and dip 60-76° towards the WNW. The dykes become increasingly thinner towards the margins. An apparently internally uniform tonalite body, 200m long and 100m wide is present 400 E. The contact relations of this tonalite and the granodiorite are unexposed.

The body at Greenauns East (G173 002) is again a dyke/host rock sheeted body. This poorly exposed body contains 1-2m wide tonalite dykes which dip 58-66° NW. The boundaries of this body are only exposed close at its NE tip where equal proportions of granodiorite and tonalite are exposed.

A 1.6km long, 210m wide tonalite body is exposed at Crillaun (G201 010). The S boundary to this body is 40m wide and contains up to 10m wide tonalite dykes intruded in Group 1 granodiorite which itself contains semi-pelite rafts up to 8m wide. The N boundary is a narrow (25m) belt of dyke/host rock sheeting which varies from 15-50cm in width. The dykes dip 68° to the NW. Granodiorite screens up to 2m wide are present throughout this body.

Towards the NE another body of tonalite is exposed at Knockaglana (G202 025). It is 1.5km long and 350m wide and consists of a swarm of tonalite dykes and host rock group 1 granodiorite in approximately equal proportions. A section through this body is exposed on the W shore of Lough Cullin (G208 028) (Fig. 3.5). Biotite granite and tonalite are juxtaposed at the northern boundary indicating that this tonalite body has utilized a pre-existing contact during intrusion. The northern boundary is a sheeted belt which has been modified by a 045° trending fault. The southern boundary is a complex sheeted belt as shown in Figure 3.5.

The tip area between the Knockaglana tonalite body and the Crillaun body contains 50cm tonalite dykes indicating that these bodies may be linked by an area of lower dilation. The sense of offset between these two bodies is again left stepping.

There are three narrow tonalite bodies exposed 400m NW from the N
Figure 3.5.

Geological map and lithology logs from the western shore of Lough Cullin (G211 030).
Fig. 3.5

LOUGH CULLIN

1a

1b

same contact 30m from 1a

continued from 1b
after 20m exposure gap

continued from 2
after 23m exposure gap

continued from 3
after 56m exposure gap

continued from 4
after 12m exposure gap

continued from 5
after 6m exposure gap

continued from 6
after 6m exposure gap

granodiorite
tonalite
granite

foliation / contact

continued from 3
after 56m exposure gap

continued from 4
after 12m exposure gap

continued from 5
after 6m exposure gap

continued from 6
after 6m exposure gap
boundary of the Knockaglana body. These bodies are 500 - 750m long and up to 80m wide and can be mapped as prominent ridges which appear on the western side of Knockaglana Hill (G202 033). They are internally homogenous and appear to have been intruded as single sheets.

3.4.3 Zone 3 tonalites

The northernmost tonalite zone is exposed discontinuously from Knockaglana Lough (G190 032) to Shannasmore (G248 057) (see Map 2). The widest part of the zone is west of Knockaglana Lough, where the tonalite is approximately 450m wide. It displays a 100m sheeted N boundary. It is offset by a 308° trending fault which has a dextral component and lies in Knockaglana Lough. The eastern side displays a more sheeted nature perhaps because a higher level is exposed, since the dyke zones are likely to be narrower and more sheeted at higher levels.

Several discontinuous tonalite bodies are exposed in the region from Pontoon (G205 071) to Cuinbeg (G229 051) At Pontoon a 200m wide tonalite body is present but the contacts are not exposed. Along the N shore of Lough Cullin there are three well exposed examples of tonalite bodies. Several 1-2m wide tonalite dykes are exposed 200m S of Pontoon Dancehall (G201 040) which have sharp contacts and provide examples of the effects that later deformation events can have on dyke contacts (see section 4.7). This tonalite dyke contains xenoliths of wall rock granodiorite and one such xenolith has a sheared, annealed margin typical of that found developed in early formed discrete shears at Pontoon Bridge (G216 049) (see section 4.2 and Plate 3.5). This provides good evidence that the tonalites where intruded into granodiorite which was at least partially crystalline. 100m to the N of the Dancehall, a 200m wide tonalite sheet with a single sharp contact is exposed. A small, late NE-SW trending fault may be responsible for the omission of the N boundary to this body.
Plate 3.5.

Small screens of granodiorite are preserved as xenoliths in the tonalite dykes. The xenolith, second from the left has a fine grained annealed margin typical of that found in the early shears in the pluton, thus indicating that the granodiorite was crystalline and deformed prior to the intrusion of the tonalite dykes. (Pontoon Dancehall, G201 039).
The Pontoon Bridge tonalite body is well exposed along the shore of Lough Cullin and in a road cutting adjacent to the bridge. Figure 3.6 shows a schematic map of the shoreline which contains the southern boundary of this body. The dykes in this boundary vary from 50cm to 25m in width and are separated by screens of group 1 biotite granite which are generally of the same width as the tonalite dykes. There are several narrow zones in this boundary in which fine scale sheeting of tonalite and granite is present. Figure 3.7 shows an example of this type of zone in which 50cm wide dykes are separated by 10 narrow screens of granite in a 4.3m wide belt. In this example they have been substantially disrupted by later deformation (see Chapter 4.).

This section also provides an example of a gently inclined contact which occurs 300m S of Pontoon Bridge where it is cross-cut by the granodiorite and tonalite foliation in a major discordant relationship (Fig 3.6). This is a result of vertical movements between the dyke walls which either fold the dyke contacts or cause intrusion into dyke walls which have themselves flat and steep sections. Detailed measurements of tonalite contact dips in the Pontoon region, show that compared to the foliation, 83% of tonalite contacts are more gently inclined than the foliation. In Figure 3.8 contact data are presented with foliation measurements in the Pontoon region for comparison. The significance of this is discussed in section 7.2. The Pontoon tonalite itself displays internal contacts between dykes of slightly differing texture due mainly to small differences in feldspar content and feldspar size in the individual tonalite dykes.

This tonalite body can be mapped for approximately 1km along strike but its northern boundary is not well exposed. There are minor tonalite units 200m N of the Pontoon body. Another body 1km east of Pontoon Bridge is 150m wide and may be related to the Pontoon Body.

Further NE another tonalite body can be mapped from Cuinbeg (G228
Figure 3.6.

Geological map of rocks exposed on the shore of Lough Cullin to the S of Pontoon Bridge (G216 049).
The tonalites exhibit faint banding and contain small alkali feldspar phenocrysts. Uniform tonalites show no banding. Granodiorite/tonalite contact is disrupted by large deformed pegmatite.

Fig. 3.6

Granodiorite, granite, tonalite, pegmatite.
Fig. 3.7

- Tonalite
- Biotite granite
- Contact dip 54-62° NW

Main foliation throughout

Lough Cullin
Figure 3.8.

Stereographic projection of structural data from the Pontoon region. Contours at > 2%, > 5%, > 10%, > 15%, > 20%.
Fig. 3.8

Pontoon poles to main foliation
n 256
mean 074/56N

Pontoon poles to sinistral shear bands
n 191
mean 046/71N

Pontoon poles to contacts
n 71
mean 073/51N
054) to Shannasmore (G299 059). There are two main exposures of this body, one close to the E shore of Lough Conn at Cuinbeg, and the other at Shannasmore. The Cuinbeg body is 150m wide and the boundaries are both narrow (125m) dyke swarms in which up to 10m dykes are present in Group 1 granodiorite. These contacts dip 48-62° NW and generally are slightly shallower than the main foliation in this area. There is a 60m exposure gap to the other exposure of this body at Shannasmore. This part of the body displays a sheeted nature throughout with up to 5m wide tonalite dykes intruded into Group 1 granodiorite.

Further NE, close to the contact at Boyhollagh (G314 106) is another large tonalite body. It is 1.5km long and 300m wide and forms a prominent hill. In contrast with many of the other tonalite bodies it is homogenous throughout. Whilst the contacts are not well exposed, no evidence was found for the existence of a sheeted boundary to this body and it is thought that its contacts are sharper. To the E this body is truncated by a major 027° trending fault which forms a large straight fault valley. The sense of movement on the fault is not known. To the south of this fault an 80m wide dyke of tonalite can be mapped which is sub-parallel to the main foliation. Sheeted contacts have not been observed at the boundaries of this body.

There is a poorly exposed tonalite body 1.5Km west of Glendaduff Lough which is 2-300m wide and up to 1Km long. There are outcrops in which a dyke/host rock sheeted zone is present but the boundaries are poorly constrained. They are parallel to the main foliation which forms in the group 1 granodiorites which are present in this area.

In summary, the tonalites are elongate zones made up of several tonalite bodies aligned along strike from each other. Each body is composed of numerous elongate tonalite sheets separated by granodiorite or granite screens. The tonalite bodies may be homogenous in their central parts. These tonalites
contain xenoliths of granodiorite which have a sheared margin which indicates that the tonalites intruded the granodiorite while it was crystalline. However, the general absence of chilled margins on the tonalite contacts indicate that the granodiorite had remained hot during this event.

3.5 The petrography of the tonalites

Petrographically the Ox Mountains tonalites can be classified as either tonalite or quartz-diorite depending on the quartz content. The tonalites from the three major zones display a similarity in their alkali feldspar content and textures in both hand specimen and thin section. They are distinguished from the diorites by the absence of hornblende. However there are intermediate diorites with biotite and hornblende, within the tonalite exposed adjacent to the large dioritic mass at Curranara (G287 030).

The Ox mountains tonalites are in general equigranular, medium grained (1-5mm) and are medium to dark grey in colour (Plate 3.7). Some variants are slightly porphyritic with up to 1cm aggregates of plagioclase and alkali feldspar. These differences can be demonstrated on the N shore of Lough Cullin close to Pontoon Bridge (G215 047) where internal dykes in the boundary zones of the tonalite sheets display slightly differing textural characteristics.

The tonalites contain white plagioclase, dark green biotite, and grey quartz with minor quantities of alkali feldspar, hornblende, chlorite and muscovite, together with the accessory minerals sphene, zircon and epidote.

Plagioclase is present as up to 5mm long elongate, subhedral to euhedral crystals which exist discretely or as components of plagioclase aggregates. Determination of composition by the Michel Levey technique indicate that the majority of crystals have a composition between An$_{25}$ to An$_{36}$. This was
supported by the Becke line test which indicated that plagioclase in these
tonalites has a higher refractive index than quartz. The plagioclase in many
cases is unzoned but both continuous normal and reverse zoning is present.
Untwinned plagioclase is common, however albite and carlsbad twinning is
present along with complex examples of twinning and zoning. Plagioclase in­
cclusions are euhedral biotite and epidote and sometimes quartz, although this
may be an exsolution feature. There is considerable development of sericite
in the plagioclase.

Biotite is generally present as elongate, subhedral aggregates 1-5mm in
length made of 0.5-1mm individual laths. The biotite may contain small
pleochroic haloes surrounding small zircons and small blebs of opaques which
are associated with biotite alteration to chlorite along 001 cleavage planes.

Quartz is present as very elongate (5mm) aggregates or 'ribbons' which
consist of 0.1-1mm, equant, anhedral crystals. The quartz crystals show evi­
dence of deformation as in most cases undulose extinction is present. There
are examples in which the quartz has completely recrystalized and undulose
extinction is absent.

Minor components in the tonalite include hornblende which are 2mm
equant, subhedral crystals which contain opaque inclusions; alkali feldspar
which is generally a perthitic orthoclase up to 4mm in size, but is not present
in all tonalite samples and anhedral muscovite which is 1mm in size and may
be an alteration product.

The accessory minerals are epidote, zircon and opaque minerals. Epidote
can be demonstrated in two situations.

(1) Euhedral 0.1mm inclusions in plagioclase phenocrysts.

(2) The epidote found outside the phenocrysts is larger, up to 1mm in
size, and it is subhedral to anhedral and forms adjacent to biotite aggregates.
Zircons are less than 0.1mm and found in biotites along with opaque blebs.
The presence of euhedral magmatic epidote as inclusions in feldspars in both the granodiorites and the tonalites has important implications. Its presence signifies that the rock was crystallizing below 8Kbars (25-30km depth)(Zen 1985, Naney 1983). The emplacement depth of the OMG is thought to be at around 24Km (6Kbars) (see section 2.5) thus implying that the epidotes are likely to have formed prior to the final ascent of the granodiorites and tonalites.

3.6 The Muscovite Granites

There are several outcrops of muscovite bearing granite which exist along the N boundary of the pluton (Map 2). These are considered to be sufficiently different from the biotite granites related to the group 1 granodiorites to warrant their division into a separate group.

3.6.1 The contact relations of the Muscovite granites

The largest area of muscovite granite stretches 3km from the east shore of Lough Conn (G223 064) to Stoneparkbrogan (G252 076). Here the muscovite granite is 500m wide and is separated from the biotite granite by a 065° trending fault which deforms a screen of metasediments exposed in a road cutting (G243 064). 500m east of Lough Conn, the northern contact with the metasediments is one in which dykes of muscovite granite up to 5m wide separate 3-5m wide metasedimentary rafts. A small 144° trending fault offsets the contact by 140m horizontally towards the NW at Stoneparkbrogan. The main contact is not exposed here but can be traced in a 5m exposure gap in two tracks at Stoneparkbrogan (G250 077) and (G254 078). Muscovite granite sheets up to 1m wide can be mapped in a 100m zone to the N of this contact.

Muscovite granite exposures have also been mapped at Carrowdoogan.
(G307 115) in two sheets which exist outside the main granodiorite pluton. These dykes are 40-50m wide and are intruded concordantly into the semi-pelites of the Attymass formation. They also occur as smaller (50cm) sheets in a semi-pelite raft at Carrowdoogan (see Fig 3.3).

Further north at Graffy (G323 132) two sheets are also present which may be the same sheet as at Carrowdoogan. Again these are located 100m outside the contact between the granodiorite and metasediments. Here the sheets are 80-100m wide with small concordant 10cm muscovite granite sheet zones in the metasediments close to their boundaries. In the townland of Ellagh Beg muscovite granite sheets have also been observed in a large metasedimentary raft at Winny Langans Lough (G344 132). In Bunnyconnellan East there are two large sheets of muscovite granite which are inside the main pluton boundary. The southern sheet (G348 149) is poorly exposed but is estimated to be 200m wide. Small semi-pelite rafts again separate this sheet from the group 1 granodiorites. The northern sheet (G365 160) displays similar properties in that it is 200m wide and has small semi-pelite rafts adjacent to its contact. This sheet can be traced north to the Gap (G372 161) where its truncated by the fault.

At Knocknasliggaun (G372 152) another major outcrop of muscovite granite occurs which is emplaced next to metasediments along its southern boundary. It is up to 700m wide and 2km long and forms a prominent ridge at Knocknasliggaun. There is evidence that this body is in a region affected by brittle faulting and the southern contact which is believed to be late oblique thrusting (see section 4.10). The northern contact between this body and the granodiorite is probably fault reactivated and outcrops of calc-silicates and psammite are present in this zone.

No evidence of a contact between muscovite granite and the granodiorite was observed so it is not possible to directly obtain the relative ages of the
muscovite granite and granodiorite, however it is considered highly probably that the presence of muscovite granite sheets in the rafts at Carrowdoogan and Ellagh beg do provide good evidence that the muscovite granites were intruded first into the metasediments of the Attymass formation and this was followed by the main granodiorite intrusion. The presence of semi-pelitic rafts along most of the contacts between the granodiorite and muscovite granite also suggests that the muscovite granite is the older. The muscovite granite has sheeted the metasediments and then as at Knocknasliggaun been enveloped by the granodiorite at a later stage.

3.6.2 The petrography of the muscovite granites

The muscovite granites (Plate 3.8) in hand specimen are quite different from the granodiorites in colour and texture. They weather to a white powdery appearance and contain obvious muscovite crystals. They are composed of alkali feldspar, plagioclase, quartz and muscovite plus accessories.

Alkali feldspar forms the majority of the feldspar component and is subhedral to euhedral, elongate and up to 5mm in size. Both microcline and orthoclase occur, although microcline predominates. Both types may be microperthitic and can display carlsbad twinning. The alkali feldspars contain inclusions of muscovite and plagioclase and are largely unaltered. Myrmekite is present at grain boundaries adjacent to plagioclase crystals.

Plagioclase is present as slightly elongate subhedral crystals up to 4mm in size. The composition averages An13 and the plagioclase appears to be unzoned. The plagioclase crystals are either untwinned or display carlsbad or albite twinning. Inclusions are mainly small (0.1mm) muscovite flakes and in rare cases alkali feldspar is present in the form of a coarse antiperthite. The plagioclase is commonly altered to sericite but this alteration is not as strong as in the granodiorites and tonalites.

Quartz is in the form of elliptical aggregates up to 5mm in length and
Plate 3.7.

Hand specimen of tonalite showing the typical dark colour, equi-granular texture and medium grain size. The mineralogy is principally plagioclase, biotite and quartz. Boyhollagh, (G311 112).

Plate 3.8.

Hand specimen of muscovite granite showing the typical low colour index, equi-granular texture and medium grain size. Principal mineral phases are alkali feldspar, quartz and muscovite with minor amounts of plagioclase. Carrowdoogan, (G307 115).
they are composed of sub-grains which are equant 0.5mm long and have wavy sub-grain boundaries. Most quartz sub-grains display undulose extinction. Quartz can form inclusions in the muscovite.

Muscovite is present as large up to 4mm, equant, euhedral and anhedral plates. There appears to be two types.

(1) One is equant and has high interference colours (3rd order)

(2) The other is anhedral with lower interference colours (2nd order); this forms laths adjacent to alkali feldspar crystals. In some cases the muscovite shows strained extinction.

Accessories include small anhedral epidote and apatite. There is a very small percentage of opaques in these rocks and biotite and chlorite are absent.

3.7 The diorites

3.7.1 The diorite contact relations

The hornblende rich diorites are confined mainly to the southern part of the pluton where they are in contact with the Group 2 megacrystic granodiorites or psammites of the adjacent country rocks.

The major outcrop of diorite inside the pluton boundaries occurs at Curranara (G287 030). This body is almost 1.7km long and 300m wide and forms a prominent ridge which is visible for several 10s of kilometres. The contact relations with the granodiorite are fundamentally different to that between granodiorite and tonalite. The diorite contacts are generally not straight, planar features (Plate 3.9, 3.10) but most are curved lobes of diorite surrounded by granodiorite. These may once have had a circular cross-section but now are elliptical due to strain imposed by the shear zone deformation. These type of contact relationships can be explained by the presence of two co-existing magmas, i.e. the host rock granodiorite was a non crystalline magma when
Plate 3.9.

Outcrop displaying the diorite contact relationships with the granodiorite. The diorite is preserved in elongate sub-rounded ellipses. These may have been pillows of diorite in the Group 2 granodiorite and suggest that the diorite was intruded into the granodiorite prior to its crystallization or that the two were co-existing liquids. Curranara (G287 030).

Plate 3.10.

Outcrop displaying diorite in contact with the Group 2 granodiorite. The lobate contact shapes suggest that they were co-existing magmas at the time of intrusion. Curranara (G286 030).
the diorite was intruded; alternatively both were intruded together. This is analagous with the 'magma mingling' of Barbarin (1988). These contact relations are common although best developed at Curranara. Identical examples have been found at Burren Hill (G264 004) in small hornblende rich enclaves which are found in granodiorite sheets intruded into psammites of the Leckee Quartzite formation.

Small intrusions of diorite occur at the margins of the tonalite bodies, e.g. at Crumlin (M173 976), 0.5km west of Lough Cullin (G202 019) and SW Lough Cullin (G223 012) (Fig. 3.4). There is another body of dioritic material which is poorly exposed 200m SW of the shore section at SW Lough Cullin. This body contains substantial quantities of biotite in association with hornblende and may perhaps represent an intermediate composition between the tonalites and the diorites. Biotite-rich diorite has also been mapped at Curranara on the margins of the diorite bodies in contact with the Group 2 granodiorites. These intermediate diorites display similar lobate contact relations suggesting that they have also been intruded into a liquid host rock.

Several large outcrops of diorite exist outside the pluton boundaries. These are confined to the metasediments to the S of the southern contact of the OMG. A major body occurs 2.5km to the S at Crumlin (M173 976) (Long & Max 1977) and it is approximately 3km long and less than 500m wide and elongate parallel to the regional strike. Granodiorite veins and sheets have also been recorded in this area by Jones (1989).

3.7.2 The petrography of the diorites

The diorite are medium-grained, 2-6mm, equigranular dark green/grey rocks comprising of hornblende, biotite, chlorite and plagioclase (Plate 3.11).

Hornblende crystals are green, equant, subhedral and up to 6mm in size and is present in two forms.

(1) As components of aggregates which form a glomeroporphyritic texture
in some of the diorite samples. These glomerocrysts are up to 6mm long and consist of individual hornblende crystals which are 0.1 - 0.5mm long. These aggregates may have formed by recrystalization following deformation (see section 6.3).

(2) A second group exhibit an equigranular texture in which single crystals of subhedral, equant hornblende show prominent exsolution along [110] cleavage planes.

The biotite content is quite variable (1-15%), but where present it takes the form of subhedral, elongate laths up to 4mm long. There is evidence that biotite is partially altered to chlorite; particularly in samples which show fine scale fracture phenomena.

Plagioclase is anhedral, up to 6mm in size, and forms a groundmass surrounding hornblende and biotite. The composition ranges from An$_{23}$ to An$_{31}$ and it is slightly zoned and displays carlsbad and albite twinning. In most cases it is highly altered to sericite.

Chlorite is generally anhedral and forms as an alteration product of biotite. Accessories include sphene and epidote, both of which can be euhedral, although anhedral epidote is more common.

3.8 Minor components

The major components of the OMG which have been described in the previous sections are in contact with various minor components such as pegmatites, aplites, microgranites, microdiorites and quartz veins.

3.8.1 Pegmatites

These have been found in all the major components. There appear to have been several different episodes of pegmatite intrusion since they display differing degrees of internal deformation. Field evidence suggests that the peg-
matites were forming more or less continuously throughout the early intrusive history. At Pontoon (G215 047) pegmatites are cut and cross-cut by granodiorite/tonalite contacts and also cross-cut by later shearing events. There are also numerous examples of pegmatites which intrude the country rocks in the vicinity of the pluton contacts and exhibit folding and boudinage. These structures form depending on the pegmatite orientation relative to the finite strain ellipsoid (see section 4.5).

The pegmatites are up to 25cm wide and composed predominantly of subhedral alkali feldspar, anhedral quartz and equant, subhedral to euhedral plagioclase. The composition is similar to the Ox Mountains granitic rocks with the exception that the pegmatites contain very little biotite. The alkali feldspar is predominantly microcline which is occasionally microperthitic. Plagioclase composition is approximately An20 and is slightly zoned and displays albite twinning. The feldspar components are generally more euhedral in pegmatites which were intruded later in the deformation history.

### 3.8.2 Aplites

Aplites occur less frequently than pegmatite veins. Their main occurrence is in the Pontoon (G205 071) region where they display irregular contacts with the biotite granite host rock. Generally they form planar sheets although some examples are irregular non-planar amorphous masses with vein patterns which radiate outwards in all directions.

The aplites are mainly composed of alkali feldspar and quartz with elongate, subhedral 0.5mm crystals of microcline and minor amounts of orthoclase. Quartz content is high (up to 50%) and is in the form of elongate, anhedral crystals which are 0.5mm long and 0.1mm wide were the aplite has been deformed. There are minor amounts of plagioclase and 0.25mm flakes of biotite are present.
3.8.3 Microgranites

Microgranite sheets are common towards the southern part of the pluton and are best exposed on the SW shore of Lough Cullin (G222 013). Here there are four parallel flat lying sheets along a 225m section (Fig. 3.4). These sheets dip gently to the S in contrast with most other contacts which are exposed in this section (Plate 3.12). There strike however, is parallel to the main foliation. The thickest sheet is 2m and the others are 1.1m, 0.5m and 0.06m wide. The microgranites contain a foliation and a stretching lineation which is parallel to the structures in the granodiorites and tonalites, thus indicating that intrusion occurred before the termination of the shear zone deformation. The microgranites intrude both tonalite and granodiorite with sharp contacts indicating both units were crystalline during the intrusive event. However, there is little evidence for a chilled margin implying that the host granodiorite and tonalite were still relatively hot during the microgranite intrusion.

The microgranites are composed of alkali feldspar, plagioclase, quartz and small amounts of biotite. Accessories include small epidote and allanite crystals. The overall composition is that of a granite. Alkali feldspar is present both as microcline and orthoclase. They are subhedral, elongate and up to 1mm in length. Both types may be microperthitic and they display carlsbad twinning in some samples.

Plagioclase is generally elongate, subhedral and up to 1mm in length and has an average composition of An15. It is zoned normally and although there is some untwinned plagioclase, in most cases albite twinning is present. Quartz forms elongate, subhedral aggregates up to 0.5mm in length which show typical wavy sub-grain boundaries and undulose extinction. There are also minor flakes of biotite present which show some alteration to chlorite.

The emplacement of these microgranites in relation to the tectonic stress fields prevalent during the later stages of intrusion is discussed in section 7.2.
Plate 3.11.

Hand specimen of diorite which is typical of those which are mainly present along the southern contact. They are usually medium grained, with an equi-granular texture. The principle mineral phase is hornblende in a matrix of plagioclase with minor amounts of biotite. This specimen shows a patch of more fine grained microdioritic material which may be xenolithic. Curranara (286 030).

Plate 3.12.

Two microgranite sheets are present in contact with Group 2 granodiorite. They are gently inclined to the South but strike parallel to main foliation in the granodiorite. The microgranites contain a weak foliation which is sub-parallel to that in the granodiorite. SW Lough Cullin (G212 011).
3.8.4 Microdiorites

There are several microdiorite bodies which have been observed on the shore section at SW Lough Cullin (G212 011) (Fig. 3.4). These have not been mapped in other regions of the pluton. They are 2m wide sheets of fine grained diorite which have steep NW dipping contacts except for one example which appears to have, at least partially a flat base. These microdiorites can contain xenoliths of a more fine grained, darker diorite (Plate 3.13). It is difficult to determine whether they have been intruded as microdiorite dykes or are present as xenolithic material which has been brought up with the tonalites and granodiorites exposed in this section.

The microdiorites are composed of plagioclase, biotite and quartz with minor amounts of hornblende. Plagioclase is elongate, subhedral and has a composition of An$_{28}$. It is normally zoned, displays albite twinning and is altered to sericite. Biotite is present as subhedral laths up to 1mm in size which form greenish brown aggregates showing some alteration to chlorite. Quartz is in the form of elongate aggregates of sub-grains which show undulose extinction. Hornblende is particularly common in the fine grained enclaves.

3.8.5 Quartz veins

In addition to the above minor igneous intrusions there are a large number of quartz veins which occur in all the components of the OMG. They exist as up to 5cm veins which are mainly related to the post-consolidation history since they cross-cut most igneous contacts. Early quartz vein emplacement can be controlled by zones of dilation in small discrete sinistral and dextral shear zones (see section 4.6 & 4.7). Late quartz veining cuts across all structures and can in some cases be related to jointing phenomena which occur throughout the pluton.
Plate 3.13

A microdiorite body in contact with the Group 2 granodiorite. The microdiorite contains xenoliths of more fine grained material. SW Lough Cullin (G212 011).
3.9 Crystallization sequences in the OMG

The OMG has undergone high shear strains during its post-consolidation history and the interpretation of any igneous features must be treated carefully. The main textural distinction which can be made is that between the Group 1 granodiorites and the Group 2 megacrystic granodiorites. The nature of the inclusions in plagioclase and alkali feldspar has been used to produce a possible crystallization sequence. There is a problem at grain margins because plagioclase and alkali feldspar may form as reaction products during the sub-solidus textural evolution of the OMG (See Chapter 6).

Crystallization sequences in the four main components are presented in Table 3.2. The presence of euhedral epidote inclusions towards the centre of plagioclase in all components of the OMG is taken as evidence that epidote crystallized first. This was followed by plagioclase, then alkali feldspar along with biotite and finally quartz. In the Group 2 granodiorites, alkali feldspar appears to have been the main phenocryst phase which strongly contrasts with the commonly held belief that it is usually last to crystallize.

Wyllie et al. (1976) reviewed the results from experimental melting of natural and synthetic granitic compositions. Their work established theoretical crystallization sequences for a biotite granite and a tonalite melt.

The sequence for a biotite granite at 6Kbars and in the presence of excess water:-

Quartz $\rightarrow$ plagioclase $\rightarrow$ biotite $\rightarrow$ orthoclase.

and for tonalite under the same experimental conditions:-

hornblende $\rightarrow$ plagioclase $\rightarrow$ biotite $\rightarrow$ quartz $\rightarrow$ orthoclase.

Naney (1983) produced experimentally derived crystallization sequences for granodiorite at 8Kbars pressure and differing water contents. The crystallization sequence for the OMG most closely agrees with that for the experi-
Table 3.2.

<table>
<thead>
<tr>
<th></th>
<th>Granodiorite</th>
<th>Tonalite</th>
<th>Biotite granite</th>
<th>Diorite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plagioclase</td>
<td>← →</td>
<td>← →</td>
<td>← →</td>
<td>← →</td>
</tr>
<tr>
<td>Alkali feldspar</td>
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<td></td>
<td>← →</td>
</tr>
<tr>
<td>Quartz</td>
<td>← →</td>
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<tr>
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</tr>
<tr>
<td>Epidote</td>
<td>← →</td>
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<td>← →</td>
</tr>
<tr>
<td>Hornblende</td>
<td>← →</td>
<td></td>
<td>← →</td>
<td>← →</td>
</tr>
</tbody>
</table>

→→→→ not developed in all samples
ment carried out at 4wt% H₂O and this implies that the OMG was intruded in an under-saturated state (Wyllie et al. 1976).

Quartz inclusions are not observed in the plagioclase crystals in the granodiorites and the biotite granites in the OMG. However, in granites in general quartz is rarely an inclusion mineral and where it is may be a sub-solidus reaction product of the feldspars and so it is not possible to comment on whether quartz was the first mineral to crystallize in the OMG. The experimental crystallization sequences of Wyllie et al. (1976) in tonalite agree with that obtained from the Ox Mountains tonalites.

The formation of alkali feldspar phenocrysts is not predicted by these models, but there is little doubt that in the OMG they are a phenocryst phase and not late porphyroblasts. They are euhedral, zoned crystals containing small inclusion of epidote and biotite close to their centres. For a review of this general problem in granitoids see discussion in Vernon (1986). The problems associated with the use of microstructural criteria for the determination of crystallization sequences have been discussed by Flood and Vernon (1988). In particular, the use of inclusions to imply the prior nucleation and cessation of growth of the included mineral is not valid if the inclusions crystallize in minute cracks in the host mineral. The entrapment of an inclusion does not necessarily imply that the included mineral has ceased to crystallize and can be explained by heterogenous nucleation of the mineral throughout the crystallization period, resulting in both inclusion and host mineral crystallizing in the melt at the same time.
3.10 Discussion of the petrography and rheology of the OMG Magmas

3.10.1 Previous work

The OMG has been previously mapped by Taylor (1966) and Crane (1984) and their petrographic subdivisions and intrusive sequences are presented in comparison with that constructed during this study (Table 3.3). Taylor’s work was to the east of the River Moy (See Map 2); he divided the pluton into three main components; granodiorite, trondjemite and basic dykes. He considered that the basic dykes were intruded prior to the main phase intrusion of the granodiorite, whereas evidence obtained during this study suggests that they are syn-magmatic (section 3.7.1).

Crane (1984) mapped the remainder of the pluton to the west of the River Moy. He divided the pluton into 14 components which were grouped in three divisions. His division 1 granodiorites are the main granodiorites which have been divided into two groups in this study (see table 3.3). He mapped the muscovite granites along the NW contact as deformed quartzites. The tonalite dyke intrusions in this study have been mapped as Division 2 and 3 by Crane (1984). The approach adopted during this study was to produce a relatively simple petrological map which would aid the structural interpretation and production of an emplacement model. The OMG is mainly composed of biotite rich granodiorite and granite which has been intruded by tonalite. Most other compositions are recognizable variations of these three rock types and can be mapped as such. An attempt to further sub-divide the main components is considered unnecessary and inappropriate.

3.10.2 Style of intrusion

Crane (1984) concluded that the internal compositional variations in the OMG could be attributed to fluctuations in diffusion rate due to changes in the
<table>
<thead>
<tr>
<th>Taylor (1966)</th>
<th>Crane (1984)</th>
<th>This study</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granodiorite</td>
<td>Division 1B-E</td>
<td>Group 1 Granodiorite</td>
</tr>
<tr>
<td></td>
<td>Division 1A</td>
<td>Group 2 Granodiorite</td>
</tr>
<tr>
<td>Trondjemite</td>
<td>Division 2 &amp; 3 granodiorites</td>
<td>Tonalite</td>
</tr>
<tr>
<td></td>
<td>(1)</td>
<td>Muscovite granite</td>
</tr>
<tr>
<td>Basic dykes</td>
<td>(2) Basic dykes</td>
<td>Appinitic diorite</td>
</tr>
</tbody>
</table>

(1) Mapped as quartzite

(2) Mapped as pre-main intrusion dykes
convection rate within the melt. He constructed a model in which at standard convection rate, fractional crystallization with intercumulus melt forms solids which are intermediate between end members of the OMG compositions. A slower convection rate would inhibit the diffusion of mafic components and the consolidation front will entrap more felsic rich solids. A faster convection rate promotes fractionation of mafic solids. The combined effect of these variations in the convection and diffusion rates produce a banded and sheeted pluton. Crane modelled the tonalites as being separate, sub-rounded bodies which had their own fractional crystallization cells.

The current study has produced a model which differs markedly from that of Crane (1984) above. The main points which are not in agreement with Crane’s observations and model are:

1. The banding produced by Crane’s model is expected to be perpendicular to the direction of maximum cooling at all times. This produces banding parallel to the walls and the roof. No evidence was encountered during this study to suggest that the banding is parallel to the roof which was mapped at Largan (G177 028). This area, although modified by brittle deformation associated with the Knockaskibbole fault system clearly shows that the roof contact is at a high angle to the steep banding and foliation.

2. Crane describes one particular type of banding which is oblique to tonalite and granodiorite contacts in the Pontoon region (G216 049) and these have been mapped in this study as discrete sinistral shear zones (see section 4.2 for details). The important point here is that banding oblique to other contacts cannot be predicted by Crane’s model.

3. Individual contacts between more felsic granodiorite and more tonalitic compositions are sharp and not gradational in any way. This implies in Crane’s model very marked variations in the convection rate of the pluton. Convection rate primarily depends on viscosity and the thickness of the body which are
not expected to vary in a rapid manner (R. England pers. comm.).

(4) The map of the tonalite bodies produced in this study (see Map 2) indicates that they are elongate bodies, not sub-rounded, as they are composed of numerous tonalite sheets suggesting that they have been intruded as dykes. The granodiorite material between the tonalites in these zones is exactly the same composition as that outside the tonalite zones. This implies that the granodiorite is not likely to have originated by a separate convection system in each body, but is part of the main granodiorite body which has dilated to permit intrusion of the tonalites.

(5) Tonalite and granodiorite banding which is greater than 10cm wide would not be expected from changes in the diffusional rate (R. Hunter pers. comm.) and the contacts would also be expected to be more gradational.

Crane’s major conclusion was that the OMG is a forcefully intruded, pre-tectonic granite which was overprinted by an amphibolite facies event. He attributed the pre-tectonic structure of the pluton to a model of fractional crystallization with intercumulus melt. This viewpoint is fundamentally different to that produced in this study which considers that the OMG is syntectonic with respect to the major D₃ sinistral transpressional shear zone. Its heterogenous internal structure is considered to be formed by intrusion of sheets of different composition into that shear zone (see Chapter 7). A fractional crystallization process may have been taking place in a magma chamber at depth with the OMG intruded by successive tapping of lower levels of that chamber. This may be where the euhedral magmatic epidotes were formed (see section 3.5) This would agree with the intrusive sequence of granite followed by granodiorite and then tonalite. There is no evidence of a strain increase towards the contacts in either the metasedimentary or granitoid rocks which would support the model of forceful intrusion advocated by Crane (see section 4.8).
The petrographic variations of the granitic rocks in the OMG is comparable with post-orogenic uplift granites classified as IKK by Pitcher (1987). The OMG is intruded prior to the late Caledonian uplift in this case, however the broad similarities with the late Caledonian granitoids can as easily be attributed to melting of the same type of lower crust. This is likely to be tonalitic or dioritic in composition and may have formed by underplating during earlier subduction of the Iapetus ocean. The OMG may be derived by partial melting of that tonalite region probably by a mantle component. Hybridization and mixing of these lower crustal melts with the mantle components may account for the diverse nature the Ox Mountains granitoids and appinitic lithologies. The future use of major and trace elements and isotopic geochemistry is likely to constrain the source region and the ascent processes for the OMG.

3.10.3 Rheology of the OMG magmas

The intrusive mechanism which has been observed in the OMG implies that the magmas where mobile during their emplacement. The magmas are not likely to have carried large percentages of crystals during their intrusion because the viscosity of such a liquid crystal mush would be too high to permit emplacement in narrow sheets or dykes. The percentage crystal to liquid ratio is likely to be lower than the critical melt percentage (CMP) which is approximately 35% (van der Molen & Paterson 1979). At percentages above this value the viscosity increases in a non-linear manner. The value of the CMP depends on a number of factors; including melt composition and temperature (Mc Birney & Murase 1984) and the distribution of melt and wetting angles in the crystal framework (Hunter 1987). Jurewicz and Watson (1984) measured wetting angles in felsic systems and found the dihedral angle to be 60°. This implies that at low melt percentages the melt will be concentrated in pools at grain edge intersections. This angle is however, reduced by several factors,
the most important of which is the water content in the granitic melt. Water saturated melts are likely to have low dihedral angles which mean that grain boundaries are likely to have a film of melt along which grain boundary sliding may occur (R Hunter pers. comm.). This means that the CMP may be as high as 98% crystal to melt ratio. Tonalitic (and probably granodioritic melts) are only saturated at 6Kbars if the water content exceeds 10% by weight (Wylie et al. 1976). The OMG is thought to have crystallized in an undersaturated state (see 3.9). This implies that the OMG magmas are likely to be 'dry' magmas and have a much lower CMP.

The OMG magmas are likely to have been intruded below the CMP and evidence from the crystallization sequences suggests that the phenocrysts present at this stage were likely to have been plagioclase and epidote. The main granodiorite magmas were intruded and on cooling, they passed through the CMP. This was then followed by intrusion of tonalite dykes into the crystallized granodiorite. The minimum time gap between the intrusion of the two units has been estimated from the relationship of Fyfe (1978):-

\[ T_{\frac{1}{2}} = 40d^2 \]  

where \( d \) is the thickness of the sheet in centimetres and \( T_{\frac{1}{2}} \) is the time in seconds to for the intrusion to cool to half the temperature difference between the wall rock temperature and the temperature of the magma during emplacement. The wall rock temperature has been estimated at 600° C (Yardley et al. 1979) and the granodiorite liquidus has been estimated at 700° C. This produces a \( T_{\frac{1}{2}} \) which will be the time for the pluton to cool 50° C. If a constant cooling rate is assumed then the granodiorite will pass through the CMP over a temperature of 50° C in an estimated time of 320 000 years. From this value the overall time to construct the OMG can be estimated at approximately 0.5Ma. Ductile shearing during this period will be interrupted
by periodic high strain rate brittle intrusion events.

In summary the OMG magmas are likely to have been intruded as hot, mobile magmas capable of being intruded as narrow sheets and dykes. This pluton cooled quite rapidly and crystallized below the CMP at which time it became further deformed by deformation in the shear zone in which had been emplaced (see Chapter 4). The overall emplacement mechanism is discussed in Chapter 7.
CHAPTER 4

THE POST-INTRUSIVE STRUCTURAL EVOLUTION

4.1 Introduction

Following the intrusion of the OMG, deformation in the sinistral shear zone continued and produced many different types of minor structures which can be related to the operation of the sinistral shear zone. It is possible to establish, using overprinting relationships, a chronological sequence of these structures (see Table 4.1 for summary). Much of the data from which the chronological sequence has been constructed comes from detailed mapping of the well exposed shore sections adjacent to Lough Cullin. Whilst the nature of the plutonic rocks exposed in these sections has been described in Chapter 3, in this chapter, the occurrence, nature and geometry of the structures is described and compared with the formation of similar structures in other parts of the pluton. These structures are described in chronological sequence in the following sections of this chapter.

4.2 Discrete sinistral shears

Discrete sinistral shears produce offsets of tonalite and granodiorite or granite. They are mainly confined to the tonalite zone boundaries where the sheeting enhances their potential for detection. They are narrow (usually 1cm wide) shears which sinistrally offset the contacts and are generally obliquely inclined to the contacts. The shears are considered to be the first structure
### Table 4.1

<table>
<thead>
<tr>
<th>Structures</th>
<th>Pontoon Bridge</th>
<th>Pontoon Dancehall</th>
<th>SW L. Cullin Bunduvowen</th>
<th>W L. Cullin Knockaglana</th>
<th>NE L. Cullin Drummin</th>
<th>S L. Cullin Muckanagh</th>
</tr>
</thead>
<tbody>
<tr>
<td>Discrete sinistral shear zones</td>
<td>3 examples</td>
<td>—</td>
<td>2 examples</td>
<td>—</td>
<td>—</td>
<td>2 examples</td>
</tr>
<tr>
<td>Main foliation</td>
<td>Strong</td>
<td>Moderate</td>
<td>Moderate</td>
<td>Strong</td>
<td>Strong</td>
<td>Moderate</td>
</tr>
<tr>
<td>Sinistral shear bands</td>
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<td>Weakly developed</td>
<td>Moderately developed</td>
<td>Moderately developed</td>
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<tr>
<td>Dextral shear zones</td>
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<td>several</td>
<td>6 examples</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>Sinistral mylonite zones</td>
<td>several examples</td>
<td>—</td>
<td>—</td>
<td>2 examples</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>Sinistral reverse shears</td>
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<td>—</td>
<td>—</td>
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<td>—</td>
</tr>
</tbody>
</table>

For location of these structures see-Map 2.
which formed in the granodiorite because the main foliation and S-C fabrics are not deflected as they cross the shears, thus implying that the main foliation development overprinted the shears at a later stage.

The formation of discrete shears prior to the main pervasive deformation in granitic rocks is unusual as it implies that a strain localization event occurred at an early (high temperature) phase in the deformation history of the OMG. Structures such as these have not been widely described in the granite literature, although similar phenomena have been noted by several authors (see discussion in 4.2.3). It is because of their exceptional development adjacent to Lough Cullin and their apparent rarity in other granitic plutons that these structures are described in a relatively detailed manner, compared with structures formed at later stages in the deformational history. Their absence in other plutons may be explained by the existence, at their time of formation, of unusual environmental conditions which produced a localized rheological response rather than the more normal pervasive fabric formation which formed later in the OMG.

4.2.1 Occurrence

The discrete sinistral shear zones are formed in three main localities on the shores of Lough Cullin and have also been observed in other parts of the pluton adjacent to the tonalite bodies.

(1) Pontoon Bridge section

There are three examples of discrete sinistral shear zones which occur in the Pontoon Bridge section (G216 049). The first is a single discrete shear which is situated 90m from the N end of the shore section (see Fig. 3.6). This shear is exposed for 20m and is straight and planar at its NE end striking 060° and dipping 46° towards the NW (Fig 4.1). This 1cm wide shear truncates 4 granodiorite screens in its footwall which are displaced 5m to the SW where they display hangingwall ramp geometry indicating a sinistral, horizon-
Figure 4.1.

Sketch of a discrete sinistral shear deforming granodiorite and tonalite. Pontoon Bridge (G216 049).
tal sense of motion. 1m to the SW the fault or shear curves into parallelism with the the granodiorite contacts in the footwall. The hangingwall granodiorite and tonalite contacts were folded in an open (3m amplitude) antiform which may have formed to accommodate displacement along the curved shear in the same way as hangingwall anticlines form adjacent to normal faults. Plate 4.1 is an oblique photograph along the probable transport azimuth of Figure 4.1

The second discrete shear zone occurs 30m S in another region of fine scale (50cm) sheeted granodiorite and tonalite. This discrete sinistral shear zone is more complex and displays a linked system containing 8 individual strands (Fig. 4.2). The shear zones dips 40-50° towards the NW and strike 030-056°. It proved impossible to detect a mineral elongation direction associated with these shear zones which could not be attributed to later deformation. The actual displacement on this system is difficult to estimate accurately due to the difficulty in matching granodiorite screens and dykes across the individual shears, however shear X (Fig. 4.2) has a horizontal displacement of 50cm with a sinistral sense of motion. The sinistral shear sense can also be reliably determined from the manner in which the granodiorite screens thin out into each shear (Plate 4.2).

The third discrete sinistral shear at Pontoon is 10m S from the linked system described above. This is a single shear which is developed mainly in tonalite, however the horizontal displacement on this shear can only be measured where it truncates a granite sheet which contains a thin tonalite dyke. This tonalite dyke is translated 4.8m horizontally on a 1cm wide shear indicating that these shears are regions of very high shear strain. The shear is extensional in nature and dips 50° NW and again the orientation of the mineral elongation direction associated with it is not known but was probably sub-horizontal.
Plate 4.1.

An oblique view, which is approximately down the transport direction of a single discrete sinistral shear. The shear dips moderately towards the NW and is a curved plane deforming granodiorite and tonalite. Granodiorite screens in the footwall are truncated, whereas they are folded in the hangingwall. Pontoon Bridge (G216 049).
Figure 4.2.

Sketch map of a linked shear system which deforms biotite granite and tonalite. Pontoon Bridge (G216 049).
Fig. 4.2

N  

0  2m  discrete sinistral shear zones

main foliation throughout

LOUGH CULLIN

Tonalite  Biotite granite  contact dip 54-62 NW
(2) SW Lough Cullin section

There are two main areas of discrete sinistral shear zones exposed at the southern end of the SW Lough Cullin section (see Figure 3.4). The first, located 70m from the southern end, consists of a linked system of shears which deform strongly banded tonalites. The banding permits the recognition of cut-offs in both the hangingwall and footwalls of the shears (Fig. 4.3). These shears dip between 44-52° NW and exhibit extensional geometries with ramp/flat segments and roll-over anticlines.

The second area containing shears is 50m further S and the shears here may also belong to a linked system, although they are slightly different to those found in the tonalites just to the N. There are four discrete shears which are planar and slightly curved in nature and are developed in granodiorite. They offset a 50cm wide tonalite dyke by 4m, 2.0m, 1.7 and 0.4m respectively (Plate 4.3). They may link just outside the present exposure (Fig. 4.4). These shears truncate the dyke contacts at slightly higher angles than those in the tonalites (22° instead of 16°).

(3) S. Lough Cullin

There are 2 examples of discrete sinistral shears along the shoreline adjacent to Friars Hill (G239 018). These offset tonalite dykes and granodiorite in a sinistral horizontal sense, although the amount of displacement is unknown. The shears are slightly curved but this is not as marked as those which deform the tonalites at SW Lough Cullin. The major difference between these shears and those described above is that here they are steeper than either the dyke contacts or the later foliation. At Pontoon or SW Lough Cullin the extensional shears are always shallower than the contacts or imposed foliation.

The shallow discrete sinistral shear zones are commonly developed throughout most parts of the pluton. They are however most easily distinguished in regions where tonalite is present, i.e. the SW part of the pluton. There are
Figure 4.3.

A sketch map of two discrete sinistral shears which deform banded tonalites. SW shore of Lough Cullin (G224 009).
Fig. 4.3

Tonalite

0 2m

hangingwall antiform

granite vein intruded along shear

incipient shear

Discrete sinistral shear
Plate 4.2.

A plan view of a discrete sinistral shear which deforms granodiorite and tonalite. The granodiorite screens are smeared and thinned out towards the shear plane which is present here along a granodiorite and tonalite contact. The manner in which the granodiorite is smeared suggests that the movement sense was sinistral. Pontoon Bridge (G216 049).

Plate 4.3.

Two sinistral shears deform a tonalite dyke which is in contact with granodiorite. One shear plane (to the left of the hammer) dips 56° towards the NW whereas the other is obscured by a set of late quartz veins (to the right of the hammer). SW Lough Cullin (G223 011).
Figure 4.4.

A sketch of three discrete sinistral shears which deform tonalite sheets in contact with granodiorite. SW shore of Lough Cullin (G224 009)
numerous examples in the Knockaglana region and on the margin of the Zone 3 Tonalite body at Gorteenamuck (G242 055)(Plate 4.4). In these examples it is rare to be able to see the actual amount of offset and determine the sense of shear, however the orientation of these structures with respect to the contacts and the observation that they are overprinted by the main foliation suggest that these structures are directly comparable with the discrete sinistral shear zones described in the above section.

4.2.2 Nature and geometry of the discrete sinistral shear zones

The discrete sinistral shears are distinguished from other structures in the OMG by the fact that they do not deflect the main foliations and S-C fabrics which are present throughout the pluton (see section 4.3 & 4.4). The shears are oblique to this foliation and are overprinted by it. The discrete sinistral shears in most parts of the pluton offset the tonalite and granodiorite contacts and are likely to have formed at an early stage after the crystallization of both components. The shears may have formed at the late stages of that crystallization process (see section 6.2). Indeed at one locality (Pontoon Dancehall, G201 040) apparently similar shears are formed along one margin of a tonalite dyke. One xenolith of granodiorite in this tonalite dyke has a margin formed from fine-grained sheared granodiorite of similar texture and microstructure to the discrete shears (see section 6.2). This implies that the discrete sinistral shears may have been forming at the last stages of intrusion in certain parts of the pluton.

The discrete sinistral shears can be regarded as extensional in nature as they produce a net extension and thinning parallel to the dyke and host rock sheeting. Whilst it is generally not possible to estimate the dip slip component in these structures. One shear at Fisherhill (221 002) shows evidence for an overthrust component to the SE (Plate 4.5) The horizontal displacement ranges from 0.5-4.8m. The structures can be described and analysed in
Plate 4.4.

A discrete sinistral shear deforms tonalite and granodiorite sheets. The shear truncates the contacts at a low angle producing a net horizontal extension of the sheeted zone. Gorteenamuck (G242 055).

Plate 4.5.

A cross sectional view of a discrete sinistral shear. The shear plane is more shallowly inclined than the main foliation. A tonalite dyke (above the hammer to the right) appears to be stretched out towards the left which would imply that the shear had an overthrust displacement component. Fisherhill (G221 012).
Fig. 4.5
Strike-slip duplex terminology

(restraining offset)  (releasing offset)  (restraining bend)  (releasing bend)

straight  separation  overlap  straight

leading extensional imbricate fan

extensional duplex

leading contractional imbricate fan

contractional duplex

trailing extensional imbricate fan

trailing contractional imbricate fan

(from Woodcock and Fischer 1986)
Fig. 4.6

Simplified sketch of linked discrete shear system at Pontoon Bridge (see Fig.4.2)

- Biotite granite

Possible reconstruction of linked discrete shear system above, demonstrating extensional strike slip duplex geometry (overall displacement not known)

- branch point
Figure 4.7.

Stereographic projection of structural data illustrating the geometry of the major fold in the granodiorite.
- tonalite contacts
- main foliation
- discrete sinistral shears

axial plane 055/60N
fold axis 06/053
interlimb angle 40°
Fig. 4.8

possible original geometry of tonalite contacts and discrete sinistral shears

This becomes folded to present orientation by gently NE plunging antiform

Fold axial plane

main foliation is axial planar
terms of the strike-slip duplex concept as proposed by Woodcock and Fischer (1986) (Fig. 4.5). The contacts between the tonalite dykes and the host rock are used as a reference plane to determine ramp and flat segments. Ramps are distinguished by the recognition of contact cutoffs in either the hangingwalls or footwalls to the shears. Both leading and trailing edge branch points between the individual imbricates can be recognised in some of the better exposed examples. Figure 4.6 is a possible reconstruction of the original geometry of the discrete shears at Pontoon (Fig. 4.2) assuming horizontal displacement only. There is no attempt to constrain the overall displacement in this duplex due to uncertainties in the displacement on individual shears and the dip-slip component. A suggested dip-slip component consisting of a limited down throw to the S would be consistent with other structures observed in the aureole and in the pluton itself (section 7.2).

The individual shears average 1cm in width and are composed of fine grained 0.1mm quartz and plagioclase feldspar with biotite flakes. The microstructure of these shears is discussed in section 6.2 however it is important to note that these are zones of extremely high shear strain with \( \gamma \) values up to 400.

The orientation of the discrete sinistral shears, dyke contacts and the main foliation are plotted in Figure 4.7. It is possible that the steep discrete sinistral shears represent the steep limb of an upright tight antiformal fold whose axis trends 055° and lies between S Lough Cullin and SW Lough Cullin, (Map 2). The axial plane to this fold is parallel to the main foliation (Fig 4.8). It is because the discrete shears display sinistral displacement on both limbs of the proposed fold that it is suggested that the original shears were formed in two conjugate sets dipping gently NW and SE which themselves deformed an original fan-like geometry of tonalite host rock contacts. This structure was subsequently steepened particularly on the southern flank of the pluton.
to form the fold structure (Fig.4.8). This fold may have formed in response to the main transpressive event which generated the main foliations and S-C fabrics in the pluton. This proposed antiformal structure and its relationship with the emplacement model for the OMG is further discussed in chapter 7.

4.2.3 Discussion

The examples of annealed granodiorite on the contacts of tonalite dykes and in rafts in the tonalite described above indicate that in certain parts of the pluton individual shears acted as planes of weakness along which late tonalite dykes were intruded. In most other parts of the pluton, however, the discrete sinistral shears clearly offset consolidated granodiorite and tonalite. In the OMG, the intrusion related structures in the wallrocks and the internal sheeting of tonalite, microgranite, pegmatites and quartz veins all provide evidence that fracture processes were in operation during the main intrusive event.

The phenomena of igneous rock fracturing at an early stage in their consolidation history has been known for some considerable time. Balk (1937) considered that the presence of fractured schlieren separated by the enclosing magma was evidence that the early crystallized parts of a magma can fracture while the remainder was still unconsolidated. The early descriptions of fractures in granitoids are classified in terms of Balks’ primary jointing structures, e.g. Mayo (1941) and Hutchinson (1956). These are often described as joints with sub-horizontal displacements and may contain granitic material, indicating they may be formed at very early stages in the crystallization history.

The formation of syn-plutonic dykes is another classic example in which granite can fracture whilst still partially molten. Pitcher (1979) quotes examples in which ‘a simple linear dyke passes along strike into an alkali feldspar-rich part of a pluton to become dismembered, appearing as lines of magma
globules'. Berger and Pitcher (1970) refer to the formation of early healed shears which provide sharp discontinuities in granite prior to its complete consolidation. Berger (1971) and Pitcher & Berger (1972) consider that the regular banding of the Main Donegal Granite was formed by the mobilization of late, potassium rich fluids into incipient early shears which formed broadly parallel to the long axis of the pluton. Berger (1971) also describes a series of early healed shears in which 'rotation and stretching of the host dark bands, a portion of which lies along the shear zone sometimes maintaining its connection and sometimes completely separate'. This may be similar to the structures which have been described above. Indeed the present author has observed evidence for sinistral offset of many banding contacts on two traverses of the Main Donegal Granite (Plate 4.6).

The formation of fracturing phenomena during the intrusion of the OMG and the subsequent development of the discrete sinistral shears are known to have occurred at amphibolite facies pressures and temperatures (see chapter 2). Amphibolite facies conditions are considered by most workers to be well below the base of the seismogenic upper crust where fractures usually form. The brittle intrusive structures formed during emplacement can be related to the influence of magmatic fluids on the shear zone stress regime (see chapter 7). The problem as to whether the discrete shears are true fracture phenomena or a type of localised ductile shearing is addressed following the description of the microstructural characteristics of these zones. (see section 6.2).

4.3 The main mineral alignment or foliation

4.3.1 Occurrence and orientation

The OMG possesses a strong foliation, labelled 'S₁', developed in all the major components and most of the minor components. The orientation is
generally NE-SW in most areas although there are some regions which display a variation from the regional trend. The foliation data is summarized on Map 2 and in Figure 4.9 a map is presented in which the pluton is divided into eight sub-areas. The foliation orientation data and means collected during detailed mapping are displayed for each sub-area.

The foliation in Cuinbeg, Pontoon and Attymass sub-areas in general demonstrates a constant dip towards the NW, however there is a 25° strike swing from Pontoon to Attymass. Further N along the NE flank the foliation dips approximately 20° more steeply than in the above sub-areas. In comparison with the sub-areas along the N contact, the SE flank sub-area displays a foliation which is steeper; the average being 75° NW with a range between 60° NW through vertical to slightly SE dipping. In general the Crillaun/Crumlin and W. Lough Cullin sub-areas display foliations with similar orientations and dips 55-70° NW. The average dip is slightly steeper than the sub-area along the N contact and there is little evidence of the foliation steepening towards the vertical adjacent to the S contact zone as in the SE flank region.

The Crillaun, W. Lough Cullin and SE flank sub-areas demonstrate a larger strike variation than other areas. In these areas the foliation locally swings from NE-SW strike to an E-W or ESE-WNW strike. It is postulated that localized zones of reverse shear, i.e. dextral shear are responsible for these narrow zones of atypical foliation strike.

The Derreens sub-area foliations demonstrate a sharp contrast to the foliation orientations in all the other sub-areas. The foliation strike is extensively modified by late brittle faulting and displays a > 90° strike variation from NW-SE through N-S to NE-SW however the foliation dip remains constant. This brittle faulting also extensively modifies and disrupts the foliation in a brittle manner and this is discussed in section 4.10. In this area the foliation
Figure 4.9.

Stereographic projection of the main foliation data and a map of the means. Contours at > 2%, > 5%, > 10%, > 15%, > 20%.
Ox Mountains poles to main foliation

Fig. 4.9

- Cuinbeg: n 89 mean 070/52N
- Attymass: n 136 mean 054/57N
- Pontoon: n 256 mean 074/65N
- Derreens: n 154 mean 013/58W
- W. L. Cullin: n 242 mean 056/68N
- Crillaun: n 197 mean 057/68N
- NE Flank: n 227 mean 060/70N
- SE Flank: n 137 mean 063/75N
dips on average 70° to the West and is parallel to the shear-zone fabric in the country rocks. Both are oblique to the gently inclined granite contact which dips 30° towards the west.

The foliation in most sub-areas is concordant with the main shear zone cleavage formed in the metasedimentary county rocks. One exception is at Ellagh Beg (G331 133) where the granodiorite foliation dips much more steeply (45°) towards the NW (see Figure 3.3). In general, the foliation dips more steeply than most internal contacts. In the Pontoon region where detailed orientation measurements have been recorded the foliation is 10-12° steeper than the tonalite contacts in that area (see Figure 3.8).

4.3.2 The mineral alignment

The foliation is a grain shape alignment fabric typically involving all or most of the phases in the granitic rock.

Biotite and quartz are the main constituents of the foliation in all hand specimens. Biotite forms very elongate aggregates of sub-grains up to 1.5cm long. Quartz also forms very elongate sub-grain aggegates up to 2.5cm long and the presence or absence of very elongate aggregates or 'ribbons' can be used to make a visual estimation of the finite strain present in each rock type.

The foliation varies in nature and intensity in the various igneous lithologies in which it is formed. This is mainly due to the differing proportions of the constituent minerals. In the granodiorites the foliation is formed mainly by alkali feldspar, biotite and quartz (Plate 4.7). The alkali feldspar megacrysts and groundmass alkali feldspar aggregates in both Group 1 and 2 granodiorites and granites are aligned parallel to the foliation. Plagioclase is generally less elongate and a preferred orientation is less discernable, but present in most hand specimens.

This mineral alignment is completely penetrative and cross-cuts all internal lithologic boundaries or contacts without any deflection. Some quartz
Plate 4.6.

A sinistral offset of banding in the Main Donegal Granite. This structure is similar in terms of its nature and geometry to those which have been studied in the Ox Mountains Granodiorite. Poisoned Glen, Main Donegal Granite.

Plate 4.7.

A hand specimen displaying the foliation which is developed in the Group 1 granodiorite which is in contact with tonalite (on the left). The main foliation is a strong, planar grain shape mineral alignment fabric. Pontoon Bridge (G216 049).
veins are not deformed by the foliation which implies that they were intruded at a very late stage in the structural history. The foliation cross-cuts the discrete sinistral shear zones (section 4.1), but is deflected by a conjugate system of late sinistral and dextral shear zones (section 4.6 & 4.7).

Figure 4.10 is a compilation of the stretching lineation data for the OMG. The lineation in general plunges gently NE or SW, except at the NE contact where it plunges gently to the west. The stretching lineation is the linear component of the $S_1$ foliation and represents the $1 + e_1$ direction of the finite strain ellipsoid; and is collinear to that developed in the country rock rafts and wall rocks. In the plutonic rocks the lineation is formed by the elongation of quartz and mica and the formation of striations which form on [001] mica cleavage planes (Plate 4.8). It is difficult to see an elongation or preferred shape orientation in either alkali feldspar or plagioclase feldspar.

The geometry of the foliation in all sub-areas is an S>L fabric as defined by Flinn (1965). The one major exception is an intense L type foliation developed at the NE end of the pluton, at the Gap (G372 161), which is thought to have been formed by superposition of a later strain component on the main foliation (see section 4.10).

Thin section examination of the foliation in all the OMG components confirms that all the constituent minerals are deformed and exhibit an alignment parallel to the planar foliation. In particular the deformation of quartz into elongate aggregates of sub-grains which show undulatory extinction in some cases implies that the OMG components were completely crystalline during the formation of the main foliation. The foliation is therefore classified as a solid state deformation type foliation. The intense nature of the solid state foliation leads to difficulty in interpreting original igneous textures and crystallization sequences and indeed the recognition of an earlier magmatic state deformation foliation. However there are certain features which may suggest
Figure 4.10.

Stereographic projection of the stretching lineation data and a map of the means. Contours at > 2%, > 5%, > 10%, > 15%, > 20%.
Ox Mountains stretching lineations

Fig. 4.10

- Pontoon
  - n 91
  - mean 13/250

- Derreens
  - n 58
  - mean 25/213

- Crillaun
  - n 82
  - mean 03/241

- Cuinbeg
  - n 82
  - mean 11/243

- Attymass
  - n 85
  - mean 02/243

- NE Flank
  - n 105
  - mean 02/241

- SE Flank
  - n 35
  - mean 05/060

- W. L. Cullin
  - n 123
  - mean 08/242
that a magmatic state fabric was present in the OMG.

(1) The recognition of a preferred shape orientation of plagioclase in most hand specimens: Plagioclase is thought to be one of the first minerals which crystallizes (see section 3.9) and will rotate in the viscous magma in response to the external stress field (Gay 1968). In parts of the pluton which show a weak foliation development it is still possible to detect the plagioclase alignment. Plagioclase behaves in a passive manner during later crystal plastic deformation (see section 6.3) and in these samples the strain intensity is not considered high enough to rotate the plagioclase. The strain is likely to be accumulated by crystal plastic deformation of more ductile minerals, such as quartz.

(2) The presence of euhedral alkali feldspar phenocrysts parallel to the foliation surrounded by mildly deformed quartz in a thin section taken from a muscovite granite sheet at Carrowdoogan (G317 127) may be a remnant of an original magmatic state foliation.

(3) Rare hand specimens of Group 2 granodiorite show evidence of tiling fabrics (section 1.4.5) in which the imbrication of alkali feldspar phenocrysts may be evidence that they rotated to touch in a viscous medium (Blumenfeld & Bouchez 1988). There is no evidence of a shear plane in these examples between the phenocrysts so they are likely to have come into contact by rotation rather than by displacement along a shear (Plate 4.9).

The possible existence of a magmatic state foliation is important because it will be the product of a syn-magmatic deformation and may provide evidence concerning the kinematics and geometry of the intrusive related deformation. In the case of the OMG the solid state deformation continued for some time after consolidation of the magma and largely obliterated the magmatic state foliation. Microstructural evidence suggests that the solid state deformation occurred close to the granitoid solidus temperatures and pressures (see section
Plate 4.8.

This outcrop displays the strong stretching lineation which is associated with the main foliation in the Ox Mountains Granodiorite. In this case it is an intense elongation of quartz which plunges 18° towards the SE indicating that the deformation was largely transcurrent in nature. Pontoon (G204 044).

Plate 4.9.

A hand specimen of Group 2 granodiorite displaying a the possible preservation of a tiling fabric. Two alkali feldspar megacrysts have been rotated into contact, thus implying that they were suspended in a viscous liquid at the time of formation. Muckanagh (G232 011).
6.2). It is considered highly probable that the shear zone deformation continued throughout the emplacement events and the early post-consolidation history of the OMG.

4.4 The S-C fabrics

4.4.1 Occurrence and orientation

A second (shear band) cleavage is widely developed throughout the OMG and this deforms the main foliation (Plate 4.10). The important geometrical and kinematic relationship between this and the main foliation, as discussed below, permit the use of the term 'shear band'. The presence of the main foliation in conjunction with the shear band cleavage is termed an S-C fabric (Lister and Snoke 1984).

In the OMG the S-C fabrics are most intensely developed along the N margin. The Pontoon sub-area in particular displays a spectacular development of such structures. In general there is a decrease in the density of the second shear band cleavage planes towards the S margin. Plates 4.11 and 4.12 display examples of the contrasting degree of S-C development at Pontoon (G215 047) and at Muckanagh on the S margin (G233 002). At the NE end of the pluton, the S-C fabrics are not well developed and in the Dereens sub-area the original sinistral S-C fabric is modified by later more brittle deformational events. The granodiorite dykes which are sheeted into metasediments along the S margin also contain the S-C fabrics which are exactly coplanar with extensional crenulation cleavages found in the adjacent metasediments. Most metasedimentary rafts in the pluton also show a concordance between their extensional crenulational cleavages and the enclosing igneous lithologies. This suggests that the formation of these structures occurred at the same time and deformed an S₃ foliation in the metasediments and the main foliation in the
Plate 4.10.

An example of the S-C fabric development which is present throughout the pluton but is particularly well developed in this region. The C planes are narrow shears formed mainly of fine grained plagioclase, quartz ribbons and elongate biotite ribbons. They deform the main foliation or S planes which is a planar grain shape alignment fabric. Pontoon (G216 046).
Plate 4.11.

A hand specimen which displays the intense S-C fabric development at Pontoon (G218 048).

Plate 4.12.

In contrast with Plate 4.11 the S-C fabrics in this hand specimen are weakly developed, however the main foliation is present. Muckanagh (G236 016).
igneous lithologies.

The orientation of the shear bands with respect to the main foliation is largely controlled by the kinematics of the shear zone system. Local variation on the scale of a thin section may be controlled by rheological factors such as the position and response of phenocrysts to the shear stress system. The shear band cleavage often forms conjugate sets which show sinistral and dextral offsets of the main foliation (see below). Sinistral shear bands strike between NNE-SSW and NE-SW and dip steeply to the NW. Figure 4.11 show orientation data plots of both sinistral and dextral shear bands in the Pontoon sub-area. Plots of contact data and the main foliation data for this region are shown for comparison. Dextral shear bands in general strike E-W to ESE-WNW and dip steeply to the N.

4.4.2 Nature and geometry of the shear bands

Individual shear bands are planar surfaces which may be up to 6cm long but are generally 0.5-2cm long. Shear band length depends on the finite shear which the sample has undergone (see Table 4.2). The spacing between shear bands decreases and therefore their density increases, with increasing shear strain as noted by Berthe' et al. (1979). Typical spacing values for the Pontoon granites and granodiorites are 0.3-0.6cm, but this is critically dependant on the grain-size of each igneous component. The tonalites at Pontoon have a shear band spacing of 0.1-0.4mm reflecting their medium grain size.

The microstructural development of the shear bands is discussed in section 6.3; briefly however, they consist of up to 1mm wide, elongate, quartz ribbons and finely recrystallized feldspar zones, with elongate biotite and chlorite present in reduced grain size aggregates.

The shear bands offset the main foliation and in many cases it is possible to restore the displacement and match up the original foliation planes. Cour-
Figure 4.11

Stereographic projection of structural data. Contours at \( > 2\% \), \( > 5\% \), \( > 10\% \), \( > 15\% \), \( > 20\% \).
Fig. 4.11

Pontoon poles to main foliation
n 256
mean 074/56N

Pontoon poles to sinistral and dextral shear bands
n 191
mean 046/71N

Pontoon poles to contacts
n 71
mean 073/51N
rioux (1984) used the angular relationship between the S and the C plane to obtain an estimate of the shear strain value. This author considers that the C planes in the OMG form as a localization of the ductile strain at a slightly later stage in the deformation history (see Chapter 6) and the angular relationship does not give an estimate of the shear strain. In this study, the average displacement on individual C planes has been used, together with an estimation of the shear band spacing, to produce a shear strain ($\gamma$) value of 0.56. This value is obtained from a total displacement of 3.4Km on the C planes and can be combined with the displacement of 16.5Km, which has been estimated from the X/Z ratios of the micro-diorite xenoliths. This is further discussed in the section on strain determination (4.8).

It is nearly always possible to determine the sense of shear on individual shear planes by examination of the relationship between the main foliation and the shear plane, both in thin section and hand specimen. The tails of the main foliation typically curve asymptotically into the C surfaces and it is the deflection of these tails which indicates the sense of shear on the shear plane.

The shear bands are clearly a product of net extension along the main foliation. They must therefore lie in the extensional field of the finite strain ellipsoid. The presence of both sinistral and dextral shear bands can be attributed to one or a combination of several processes.

(1) A coaxial component producing flattening perpendicular to the main foliation will generate both sinistral and dextral shear bands which are symmetric to the main fabric (Fig. 4.12a).

(2) The partitioning of strain into narrow shear bands in a non coaxial regime may cause the formation of synthetic and antithetic shear bands which are asymmetric to the main fabric. The synthetic shear bands are generally formed parallel to or at a low angle to the bounding shear couple and the antithetic shear bands are formed at a higher angles The latter may rotate to
become inactive (Berthe' et al. 1979, Platt and Vissers 1980) (Fig. 4.12b).

It is not possible to quantitatively determine the magnitude of the coaxial component during deformation in the OMG, but it is expected to be a significant component in a transpressional shear zone.

The overall sense of shear represented by these fabrics can be reliably determined visually by recognising that:-

(1) The number of sinistral shear bands developed is at least an order of magnitude greater than that of the dextral shear bands.

(2) The sinistral shear bands are visually more apparent than the dextral shear bands due to the greater development of fine grained polymineralic aggregates in the former.

(3) The displacement on the sinistral shear bands is much larger than that on the dextral shear bands.

4.4.3 Discussion

S-C fabrics were first described by Berthe' et al. (1979) in granitic rocks and orthogneisses deformed in the South Armorican Shear Zone. He originally proposed that the angle between the shear plane and the main foliation decreases with increasing shear strain. Table 4.2 shows angles which have been measured for samples from the OMG. These are from rocks which display a relative strain gradient. The data shows a positive correlation with with the predictions of Berthe' et al. (1979). As strain qualitatively and quantitatively increases (xenolith X/Z ratios):-

(1) The S-C angle decreases.

(2) The spacing between the C planes decreases.

(3) The C plane length increases.

It would seem from these data that very large values of shear strain are necessary to rotate the main foliation parallel to the C surfaces. However individual hand specimens may show up to 10° variation in the angle between
Fig. 4.12

(a) coaxial deformation

(b) non-coaxial deformation

after Platt & Vissers (1980)
<table>
<thead>
<tr>
<th>Sample</th>
<th>MN21</th>
<th>PD12</th>
<th>P51</th>
<th>A45</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>Muckanagh</td>
<td>Pontoon</td>
<td>Pontoon</td>
<td>Attymass</td>
</tr>
<tr>
<td>GR</td>
<td>233 002</td>
<td>205 071</td>
<td>215 047</td>
<td>307 115</td>
</tr>
<tr>
<td>Fabrics</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>S-C dev.</td>
<td>weak</td>
<td>moderate</td>
<td>mod/strong</td>
<td>v. strong</td>
</tr>
<tr>
<td>S-C angle</td>
<td>35°</td>
<td>29°</td>
<td>27°</td>
<td>13°</td>
</tr>
<tr>
<td>C spacing</td>
<td>0.4mm</td>
<td>0.4mm¹</td>
<td>0.3mm</td>
<td>0.2mm</td>
</tr>
<tr>
<td>C length</td>
<td>2mm</td>
<td>3-4mm</td>
<td>6mm</td>
<td>20-30mm</td>
</tr>
<tr>
<td>Strain</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Fry inclin.</td>
<td>-</td>
<td>+18°</td>
<td>+38°</td>
<td>+64°</td>
</tr>
<tr>
<td>Fry X-Z ratio</td>
<td>-</td>
<td>2.4</td>
<td>2.1</td>
<td>2.8</td>
</tr>
<tr>
<td>Xen X-Z ratio</td>
<td>8.0</td>
<td>9.5</td>
<td>14.8</td>
<td>______²</td>
</tr>
</tbody>
</table>

¹ PD12 is a medium grained tonalite so has closer spaced C planes compared with the granodiorites.

² Xenoliths have not been measured adjacent to this locality however mylonites with similar fabrics at Bunduvowen contain xenoliths with X/Z shape ratios > 20.

Table 4.2
the C surfaces and the main foliation. This angle greatly depends on the presence or absence of feldspar phenocrysts in the fabric. Some hand specimens also show evidence for the development of more than one set of shear bands and the second set is usually at a higher angle to the foliation than the first set. Estimates of strain by measuring the angle between main foliation and individual shear bands is not likely to yield a meaningful result as the shear planes form late in the OMG, furthermore the original angle between the two planes is not known. In the country rocks extensional crenulation cleavages, which are thought to be of a similar age as the S-C fabrics, make angle of 30° to the main fabric and this does not vary greatly with the strain gradient (C.S. Jones pers. comm.). In the OMG the angle remains at approximately 35-45° to the foliation until large scale recrystallization and grain size reduction occurs which is mainly in localized zones in the OMG. In these zones the angle may be reduced to 20-25°.

Berthe 'et al. (1979) considered that both planes formed more or less simultaneously to produce the S-C fabric at an early stage in the deformation history. In the OMG, the presence of areas where the shear bands are not strongly developed, despite the main fabric being quite well formed suggest however that the C planes develop at a later stage in the deformational history. This may be controlled by a change in feldspar deformation mechanism which occurs as the pluton cools (see section 6.3) and it is likely following Platt and Vissers (1980) that that the shear bands may form later due to late stage strain partitioning and localization during the non coaxial deformation.

4.5 Folding in the OMG

Various types of folds are developed in the OMG. A major fold has been inferred from the change in vergence relationships between the discrete sinis-
tral shears and the contacts, as described in section 4.2. This fold is likely to be upright and the main foliation which dips steeply NW is axial planar to it. The wavelength and amplitude of this fold is not known. It is likely to have developed at an early stage following the intrusion and may have been forming at the same time as the discrete sinistral shears. This fold is asymmetric with a southern limb which is vertical or slightly overturned and a moderately NW dipping northern limb. It is possible that this fold corresponds to the major F$_3$ antiformal structure developed in the metasediments (see section 2.) whose axial trace lies along the area now occupied by the OMG.

There are numerous examples of minor folds developed in the OMG which deform the tonalite and granodiorite sheets, pegmatites and quartz veins. These are mainly upright, open to close with axial traces, aligned NE-SW. At S. Lough Cullin (G212 011) a fine scale tonalite and granodiorite multilayer is folded to form 2 asymmetric fold pairs with a wavelength of 50cm and amplitude of 1m (Plate 4.13). The fold axis is parallel to the stretching lineation but the axial planes have a slightly shallower dip to the NW than the main foliation developed in the adjacent tonalite. The main foliation appears to have been folded by these folds which indicates that they formed at a late stage in the deformation chronology. The fold limbs have been attenuated by discrete sinistral shear. Geometrically these folds are a product of NW-SE directed shortening. Whilst the sense of vergence is not correct for sinistral shear, the fold limbs have been significantly modified by sinistral shear to change the original vergence. These folds are likely to have formed at a late stage by localized shearing between adjacent tonalite and granodiorite sheets.

Pegmatites and quartz veins form upright, open to close folds with axial planes parallel to the main foliation. Their geometry suggests that they formed in response to NW-SE directed shortening formed late in the evolution of the shear zone. At Pontoon Dancehall (G201 040) a 5cm wide pegmatite vein is
folded to form an open, upright fold (Plate 4.14). Minor granitic veins are folded at Pontoon Bridge (G216 046) and have attenuated limbs. Their fold axes are parallel to the stretching lineation which plunges gently SE in this section.

Other folds in the pluton can clearly be related to the early discrete sinistral shearing (see section 4.2) and the later dextral shear zones described in the next section. The early discrete shears produce open folds in their hanging walls which is an accommodation feature to the curved listric shear profile. Late dextral shears produce sigmoidal folding of the main foliation because they form at 25-40° to the main foliation. The foliation is reorientated towards parallelism with the dextral shear by drag on the shear plane.

4.6 Late sinistral shear zones

Late sinistral shear zones are variably developed throughout the pluton. They are termed 'late sinistral shear zones' and distinguished from the early sinistral shears because they deform and intensify the main foliation and S-C fabrics into localized zones of mylonite.

They are distinguished in hand specimen and outcrop because they form narrow bands of fine grained aggregate which are darker than the surrounding less deformed granitoids. The best examples are located in outcrops along the shores of S. Lough Cullin. Late sinistral shears are mainly developed in the Pontoon region which may have undergone higher values of bulk strain than other regions of the pluton (see section 4.8). These late mylonite zones are widely developed 300m S of Pontoon Bridge (G216 046)(Fig. 4.13) where they are up to 3cm wide and sub-parallel to the main foliation. There are numerous late mylonite zones here in a 50m wide belt (across strike) which all cut across the foliation at low angles. Significantly they cut across the foliation in the
Plate 4.13.

Folding of fine scale granodiorite and tonalite sheeting in which the fold limbs are modified by sinistral shearing. SW Lough Cullin (G223 010).

Plate 4.14.

An upright fold of a pegmatite vein. The main foliation is axial planar to this fold and its axis is sub-parallel to the stretching lineation. Pontoon (G218 047).
Figure 4.13.

Sinistral mylonite zones deform biotite granite and tonalite 500m to the S of Pontoon Bridge (G216 049).
Fig. 4.13

- Biotite granite
- Granodiorite
- Mylonite zones
- Strike-slip duplexes
- Less deformed granodiorite
opposite sense to the early discrete sinistral shears which implies that they are strike-slip compressional structures as opposed to the extensional discrete sinistral shears.

This outcrop contains excellent examples of strike-slip duplexes (Woodcock & Fischer 1986) developed in the mylonites and these shears divide the outcrop into large blocks of granite, granodiorite and tonalite (Plates 4.15 & 4.16). This duplex can be classified as a contractional strike-slip duplex because there is contraction in the hangingwall. The transport direction, as indicated by the stretching lineation is 08° to 242° suggests that the displacement was predominantly transcurrent. Duplexes are developed on two scales; small scale, 20cm long imbricate slices and larger, more common 15m imbricates which divide the outcrop into large blocks. The smaller scale duplexes are developed close to the highly deformed contact between the granite and the tonalite. This contact has localized the late ductile deformation producing a zone of imbricate slices or blocks of less deformed granite and tonalite, bounded by mylonites, which have been detached and transported along the former contact or detachment horizon. Individual duplexes show normal duplex geometries. Thus leading and trailing edge branch points and roof and floor shears can be identified allowing the use of terminology which is normally applied to deformed sedimentary and metasedimentary rocks. A general lack of marker horizons makes determination of displacements difficult.

Other examples can be found at N. Lough Cullin (Fig 4.14), where several screens of granodiorite are enclosed by a tonalite dyke and these originally separated blocks have been moved around and brought into contact by movement on mylonite zones. These shears are extensional in nature and contrast with those above (Fig. 4.13). This outcrop also displays, in a less spectacular manner, that the contact can become a rheological boundary along which deformation may localize and along which screens can be translated so as to
Plate 4.15.

Large scale mylonite development along a contact between biotite granite and granodiorite. The mylonite geometries resemble strike-slip duplexes and translate imbricates of biotite granite and granodiorite along the contact. Pontoon Bridge (G216 046).

Plate 4.16.

A small scale example of a strike-slip duplex in which three imbricates are stacked on top of each other and are individually surrounded by narrow mylonite zones. Pontoon Bridge (G216 046).
Figure 4.14.

Sinistral mylonite zones deform granodiorite and tonalite on the northern shore of Lough Cullin (G334 047).
Fig. 4.14

folded granodiorite mylonite truncated by tonalite contacts

3-4 cm tonalite vein

shear disappears into zone of distributed higher strain

late sinistral shear zones

granodiorite

tonalite
bear little resemblance to the original intrusive geometry. (Plate 4.17)

The shore section 200m S of Pontoon Dancehall (G211 039) contains examples of sinistral shears (Fig 4.15). The dyke contacts are offset up to 75cm along fine grained < 1 cm wide shears. These shears are preserved at a relatively high angle to the dyke contacts and the main foliation (maximum 47°) and can be traced for a small distance into the granite host rock where they disappear into regions of distributed strain. These zones represent imbricate tips and are considered to have developed during the early stages of the duplex formation, before deformation ceased in this part of the pluton. Granite and tonalite dykes and a folded pegmatite vein are cut off in the footwall by a shear which is parallel to a contact; in contrast the other shears cut across the granite and tonalite contacts. This imbricate system shows a slight back-steepening as structurally higher shears are inclined at higher angles to the main foliation. This system can also be classified as a contractional system in a similar manner to that at Pontoon Bridge in which case there is likely to be a thrust component along these shears.

This author has observed similar phenomena in dykes in the Dinkey Creek pluton in the Sierra Nevada and in a photograph from the Flamanville granite pluton in Hutchison (1956). The ‘A joints’ proposed by Hutchison displaced an incongruent vein without rupture and this produced offsets of vein walls which appear remarkably similar to those at Pontoon Dancehall. At Pontoon Dancehall the tonalite dykes did rupture to produce thin comminuted zones along the shears.

The geometry of the sinistral strike-slip duplexes is described below following the description of similar dextral shear zones which display evidence that they may be a coeval structure.
Narrow mylonite zones which develop as a localization of the later stages of the shear zone deformation on the contact between the biotite granite and the granodiorite. The biotite sheets are brought into contact by movement on the mylonitic zones. Pontoon (G218 048).
Figure 4.15.

Sinistral reverse shears deform tonalite dykes at Pontoon Dancehall (G212 039).
Fig. 4.15

- granite
- tonalite
- pegmatite

**Legend:**
- Dextral shear zone
- Xenolith
- Tip zones of more homogenous strain
- Sinistral mylonites
- Early sheared granodiorite on one contact

**Note:**
- Scale: 0 1 2 m
- Orientation: North (N)

Locations:
- Lough Cullin
- Cullin granodiorite
4.7 Dextral shear zones

4.7.1 Occurrence and orientation

Dextral shear zones occur commonly throughout the OMG. The best exposed examples occur along the shores of Lough Cullin and several examples are described from these sections.

(1) Pontoon bridge section

A dextral shear zone system occurs 70m S of the Pontoon Bridge (G215 049) (Fig 3.8) and here there are two closely spaced dextral shears which deform the main foliations, S-C fabrics and three screens of granite (Fig 4.16). The granite screens are deformed by drag along the principal dextral shear to produce open folds. The central granite sheet is detached from the shear plane creating a small cavity into which late quartz rich fluids have been emplaced. Small dextral shear bands are formed in the granodiorite adjacent to the shear zone. The other shear zone is developed 1m to the S and does not produce a discrete displacement; however it does produce open folding of the main foliation and sinistral shear fabrics. The principal dextral shear is cut and displaced by a late sinistral shear zone which in turn is folded by the minor dextral shear zone indicating that both sinistral and dextral movements were contemporaneous.

(2) Pontoon Dancehall section

There are numerous examples of dextral shear zones along the section 200m S of Pontoon Dancehall (G211 039). Plates 4.18 and 4.19 are examples in which a pegmatite and a xenolith are deformed by the late dextral shear zone. In this section there is also evidence for the dextral shear zone producing more profound reorientation of tonalite dyke contacts. Figure 4.17 is a field sketch of the tonalite contacts along this section which have been deformed by open folds in response to localized dextral shearing. This dextral shearing
Figure 4.16.

Late dextral and sinistral shearing deform tonalite and granodiorite contacts at Pontoon Bridge (G216 049).
Fig. 4.16

LATE SHEARS DEFORMING SHEETED CONTACTS AT PONTOON BRIDGE

KEY

Granodiorite

Tonalite

Igneous contact

Foliation

Discrete shear

Granodiorite

Tonalite

Quartz in releasing band

Dextral shear bands locally developed

2 cm ultra-mylonite

N 60° main foliation

0 1 m
Plate 4.18.

A microdioritic xenolith which is deformed in a late dextral shear zone. Movement on this dextral shear has created an area of dilation in which quartz has been emplaced and deformed. Pontoon Dancehall (G211 038).

Plate 4.19.

A late pegmatite vein is displaced along a dextral shear. Pontoon Dancehall (G211 039).
Figure 4.17.

Late dextral shear zone deforming tonalite and biotite granite contacts at Pontoon Dancehall (G212 039).
Fig. 4.17

- Granodiorite
- Tonalite
- Pegmatite

Tonalite dyke folded by dextral shear zone

Ox mountains granodiorite
also deforms the main foliation and sinistral S-C fabrics and is taken to have formed slightly later than them. Further along the shore line at (G211 038) there are three dextral shears which also produce open folding of the main foliation.

(3) SW Lough Cullin

There are 6 examples of dextral shear zones exposed on the shore line at SW Lough Cullin (Fig 3.4). They are located in a 70m section which is 60m from the N end of the shore section (G011 223). The dextral shear zones deform tonalite and granodiorite contacts and microdiorite and granodiorite contacts, as well as the main foliation and the weakly developed S-C fabrics.

In all areas in which they have been observed, the dextral shear zones display a constant orientation with respect to the main foliation and contacts. Figure 4.18 is a plot of the orientation of dextral shear zones formed in the OMG. In general they dip steeply to the N or NNE.

4.7.2 The nature and geometry of the late dextral shear zones

The dextral shear zones are generally recognised because they fold the main fabrics and internal granite contacts causing obvious dextral deflections of contacts in particular. The width of these zones is variable; each individual shear zone may be up to 20cm wide but may contain a central zone of highly comminuted mylonitic material. The median zones are up to 2cm wide and composed of a fine grained, polymineralic aggregate of quartz, feldspar and minute biotite flakes. The boundaries to these median zones may be sharp but are more commonly gradational into less deformed granitoid which displays evidence of dextral shear band development, overprinting the sinistral shear bands.

The formation of the dextral shear zones can be attributed to one of two possibilities.

(1) A separate, later dextral shear couple deformed the earlier formed
Figure 4.18.

Stereographic projection of poles to dextral shear zones. Contours at > 2%, > 5%, > 10%, > 15%, > 20%.
Fig. 4.18

Ox Mountains poles to dextral shear zones

n 34

mean 106/84N
sinistral structures.

(2) The dextral shear zones are formed in zones of antithetic shear to the main sinistral shear zone, which at this stage had become increasingly partitioned into localized mylonite zones.

The second interpretation is supported by the observation that both sinistral and dextral shear zones deform each other at Pontoon bridge (see above).

The geometry of the late dextral and sinistral shear zones relative to the shear zone is discussed in section 4.11.

4.8 Strain measurements in the Ox Mountains

4.8.1 Results from xenolith measurements

The methods used for the measurement of strain in granitic rocks have been briefly described (section 1.3.4). Hutton (1982, 1988b) used the shape ratios of microdiorite xenoliths in the Main Donegal Granite and Strontian Granite to estimate the relative variation in the shape of the strain ellipsoid. In the OMG xenolith shape measurements were made where possible during this study. The nature of the exposures, which consist mainly of gently inclined, glaciated surfaces exposed in drift, mean that it is rarely possible to measure the three dimensional shape of the strain ellipsoid and most of the data have been recorded on horizontal surfaces which are considered to contain the $X/Z$ section of the finite strain ellipsoid. The data presented in Figure 4.19 consist mainly of $X/Z$ ratios with the $K$ value where this has been determined.

The data are not sufficiently numerous to provide detailed strain profiles across the pluton. This is due mainly to the scarcity of microdiorite xenoliths in the OMG. The errors inherent in these data are likely to be statistically large due to the low numbers of enclaves which some of these averages represent. Populations varied between 5-30 per locality with the average being 12. This
Fig. 4.19

average X/Z shape ratios — 12.2

K value — (0.43)
may appear to question the validity of the apparent relative strain variations, however it should be noted that such variations were always accompanied by an observable variation in the fabric intensity.

A composite strain profile across the pluton is shown in Figure 4.20. This shows that there are at least two high strain zones within the OMG. These zones may correspond with the zones of tonalite intrusion. A high strain area is formed to the south of the Zone 1 tonalite body. It is aligned parallel to the tonalite body, both of which are oblique to the southern contact of the main pluton with the metasediments. A high strain region also exists between the Zone 2 and Zone 3 tonalites. There is no xenolith data to confirm the presence of a high strain zone, but qualitative estimation of fabric intensity in this region suggests that there is a high strain zone along the margin of the Zone 1 tonalite body.

The Zone 3 tonalite body may exhibit a high strain zone which can be observed in the Pontoon sub-area where it is represented by the development of high strain mylonites as the strain becomes localized during the late stages of the ductile deformation.

The Curranara region to the south of Foxford displays moderate strain intensity, as does the small granodiorite pod at Burren hill. The high strain zones in the OMG are located inside the pluton, the exception being along the northern contact of the highly deformed muscovite granite at Stoneparkbrogan which may be due to later movements along the Knockaskibbole Fault system (see section 4.10). The fact that the pluton contacts are not highly strained has important implications for the emplacement mechanism, indicating that pluton was not forcefully emplaced. This is further discussed in Chapter 7.

The average $X/Z$ shape ratio (10.5) measured in the Ox Mountains is in general high when compared with most other plutons. The amount of horizontal displacement which this represents, assuming simple shear ($\gamma =$
mean \(X/Z\) shape ratio

Fig. 4.20

Composite strain profile through the OMG
2.7), has been estimated at 16.5Km. This produces a total of 19.9Km when added to that estimated for the S-C fabrics (see section 4.4). This value is for the displacement present in the OMG alone and ignores the considerable displacement which occurred in the metasediments prior to intrusion and the later history of localized shearing in the OMG. This must be counterbalanced however by the fact that the xenolith based shear strain determinations are likely to be a maximum since:-

(1) The assumption of simple shear ignores the pure shear component which is likely to be significantly large in a transpressional shear regime.

(2) The X/Z ratios may overestimate the ellipticity of the strain ellipsoid because the micro-diorite xenoliths where more ductile compared to the surrounding granitoid. Their finer grain size may have contributed to this over estimation by localizing deformation.

### 4.8.2 Results from the Fry analysis method

The spatial distribution of feldspar phenocrysts in a granitoid which has undergone deformation will be non-uniform (Ramsay 1967, Fry 1979). The Fry method assumes that the centres of phenocrysts were initially an isotropic, non-Poisson distribution. The centres will rotate passively and will assume a non-uniform distribution which will reflect the shape of the strain ellipsoid. Salter and Hodges (1988) proposed that the sense of shear in non-coaxial deformation may be inferred by measuring the degree of asymmetry of the strain ellipse axis to the foliation plane.

A trial analysis was carried out on hand-specimens to assess the usefulness of the method for a more widespread study. Initial trials using enlarged photographs of thin sections proved that most lithologies were too coarse grained and too few phenocrysts were present to define an ellipse properly. Analysis were performed on slabs of OMG granitic rocks cut perpendicular to the foliation and parallel to the lineation (XZ plane). The method used is that
proposed by Hannah and Fry (1979) and Ramsay and Huber (1983), in which the positions of all phenocrysts are marked on a clear acetate sheet which contains parallel reference lines. A second sheet with a single centre point and parallel reference lines is placed so that the centre point corresponds to the position of a phenocryst on the first sheet. The position of all other phenocrysts are marked on the second sheet ensuring that the reference lines remain parallel. Fry (1979) states that it is important to mark all other points rather than just the nearest neighbours because of the importance of the initial distribution of the phenocrysts. A truly random, or Poisson, distribution will not produce an ellipse after deformation, whereas an initially homogenous or isotropic distribution will. The use of all phenocrysts or points in the distribution, although more time consuming if applied manually, provides better definition of the ellipse.

Figure 4.21 provides the results from the preliminary study. The ellipses are shown with reference to the lineation (S) and shear plane (C) in a similar way to that shown by Salter and Hodges (1988). The angle of inclination is taken as positive moving in a clockwise direction from the C plane. The results from this study show that the angle of inclination varies from -14 to +85. Salter and Hodges (1988) proposed that positive inclinations from the C planes indicate sinistral sense of shear whereas negative inclination indicate an antithetic dextral sense of shear. This author does not agree that a negative inclination necessarily indicates dextral sense of shear and it is considered that positive angles up to 45° indicate sinistral shear and angles over 45° should indicate dextral shear (Fig 4.22). It is not precisely certain what a negative inclination implies, if anything, about the sense of shear. The wide range in angles obtained from the Ox Mountains samples indicate that it is difficult to determine accurately the sense of shear from the orientation of the ellipse. Samples which display well developed conjugate shear band cleavages tend
RESULTS FROM FRY ANALYSIS OF 14 OMG SAMPLES

Dotted line outlines a secondary ellipse

Key
- sample number
- grid reference number
- inclination from shear plane
- X/Z shape ratio

G/d1 G227 055
+38° 2.5

G/d2 G216 045
+64° 2.6

G/d3 G215 045
+80° 4.1

G/d4 G325 103
+67° 3.2

G/d5 G345 114
+24° 1.7

G/d6 G306 042
+85° 2.8

G/d7 G214 054
+28° 3.0

G1 G370 151
+70° 4.2

G2 G314 114
-19° 2.5

G3 G317 127
+45° 6.1

Diorite G280 028
-72° 2.3

Ton1 G324 113
+31° 14

Ton2 G212 038
+18° 2.5

Ton3 G216 046
-14° 6.1
Salter & Hodges (1988) suggest that a negative inclination indicates dextral shear, however this implies that the strain ellipse is $45^{-90^\circ}$ to foliation.

This study proposes that angles up to $+45^\circ$ indicate sinistral shear and angles between $+45^\circ - 90^\circ$ are antithetic.
to produce inclinations > 45° whereas predominantly sinistral shear banded samples produce inclinations < 45°.

The XZ ratios which were obtained from the ellipses varied from 1.5-12.8, however most samples have a ratio in the 2.1-4.5 range. The higher ratio results came from samples which look less deformed, i.e. little shear band development. In particular samples G/d 1,2,5,7,Ton2 and Gr2 have ratios under 3.0 and these samples in hand specimen have well developed shear band fabrics (see Plate 4.11 for example). G/d 2 in particular is a highly strained mylonite which shows a very strong grain size reduction in shear planes surrounding relict feldspar phenocrysts (Plate 4.21).

In general the application of the Fry technique was not successful for the following reasons:-

(1) Poor definition of the ellipse was caused by the low numbers of points used in each analysis. The maximum used was 116 whereas Fry (1979) recommends 300.

(2) The presence of two ellipses in some cases leads to confusion over which is the real ellipse. This was due to either too few points or to some undefined function of the conjugate shear bands in some samples.

(3) The angle of inclination of the ellipse can be highly variable due to the presence of conjugate cleavages in some samples.

(4) The OMG rocks are products of high levels of shear strain which is seriously underestimated by the Fry analysis particularly those which show heterogenous strain, i.e. shear bands and mylonite development. Fry (1979) recommends that the method should be used on samples with X/Z ratios less than 6.0 and it is likely to be more useful for granitoids with relatively low strain ratios.

(5) The grain size of certain granitoids created some difficulties. Medium-grained tonalites are too fine grained and produce a small ellipse and they
produce better results using enlarged photographs of thin sections. Coarse-grained granites are too coarse for this method as they contain too few phenocrysts in a thin section. Even in hand specimen the ellipse produced may still be quite small. The answer may lie with the use of enlarged photographs of rock slabs.

As previously indicated, the Fry method may seriously underestimate the ellipticity of the strain ellipse. In general, for rocks which are as deformed as the OMG, it is necessary to use xenolith shape ratios as in section 4.8.1 for strain estimation. These data may however overestimate the strain due to the presence of a ductility contrast between granodiorite and the mafic xenoliths. The OMG xenoliths show a fine grained-texture and may contain a more highly developed fabric than the granodiorites indicating that they were more ductile than the surrounding granodiorite. This should not affect the relative changes in strain intensity across the pluton.

4.9 Qualitative fabric intensity

In order to provide a representation of the degree of foliation and S-C fabric development, data presented from four samples (Plates 4.11, 4.12, 4.20 & 4.21) which provide examples of the differing degree of fabric intensity and therefore finite strain.

Table 4.2 displays a comparison of various properties of each sample and provides an illustration of how a qualitative scale of fabric intensity can be used in the field. Results from the Fry analysis and X/Z ratios from xenoliths are also presented for comparison. As previously noted, the Fry method does not appear to provide useful results for the higher strain rocks which were deformed heterogeneously. Xenolith measurements are likely to be more accurate but highly dependant on their occurrence throughout the pluton. In
Plate 4.20.

A hand specimen of tonalite with a well developed foliation and poorly developed shear bands. Pontoon Dancehall (G211 038).

Plate 4.21.

A hand specimen which displays the typical fabric development in the late sinistral mylonite zones. Attymass (G301 095).
the Ox mountains a scale such as that presented in Table 4.2 can be used with practice to define regions of high, low and medium strain.

4.10 Late faulting in the Ox Mountains

There are several late faults which deform the Ox Mountains Granodiorite. The most important of these is the Knockaskibbole Fault (Long & Max 1977) which is located between Coarsepark (M139 919) and Lough Conn (G190 056). This was originally identified by Currall (1963), who described it as a major thrust. He identified a Sole thrust which he mapped on the shore of Lough Conn and a structurally higher Conloon thrust. These two structures isolated a wedge of the granodiorite (Dereens sub-area in this study) in which highly crushed and cataclastic granodiorite was observed, formed by SE directed thrusting. Long and Max (1977) noted that the Knockaskibbole Fault had a dextral displacement of approximately 10Km. This study indicates that the Knockaskibbole Fault has a protracted history, in which early dextral semi-brittle fabrics deform the main OMG sinistral fabrics in the Dereens sub-area. The work of Jones (1989) has shown that this is consistent with 10Km NE displacement of the Lough Talt and Glennawoo shear zones by dextral movement on the Knockaskibbole Fault (See Map 1).

Long and Max (1977) suggest that the Knockaskibbole Fault is a splay of the North Ox Mountains Fault, which is coincident with the Fair Head - Clew Bay line of Max & Riddihough (1975). They provide evidence for syn- to post-Carboniferous movement on this fault system. Deformation of this age is likely to be preserved in the late, steep breccia zones which cut the region of dextral fabrics at Lough Conn. Here, there is a series of late N thrusting imbricates which are cemented with calcite perhaps suggesting a Carboniferous dextral movement on the Knockaskibbole Fault (Plate 4.22).
Plate 4.22.

Late brittle deformation which deforms the Ox Mountains Granodiorite at the SW end and is associated with dextral movements on the Knockaskibbole Fault system. Here a series of North thrusting imbricates brecciate the granodiorite. These are formed on a restraining bend on the dextral fault zone. Terrybaun (G190 056).
The other major fault is NW-SE trending, passing through The Gap 500m to the north of Lough Talt (G405 147). Here the fault truncates the OMG, but the exact movement sense is not known. This structure may be related to S₃ fabric swings from NE-SW to NW-SE at Attymass (G295 115) and Benalta (G390 170) (see Map 1). These structures have also deformed the NE end of the pluton at The Gap, creating intense I type fabrics by superposition of a flat lying E directed thrust fabric on an earlier steep fabric. The exact nature and cause of these late structures remain unclear.

Small faults at Curranara (G287 030) and Stoneparkbrogan (G252 076) displace the main pluton contact 500m NNE. These faults are not however, exposed.

4.11 Summary and discussion

The discrete sinistral shear zones, main foliation, S-C fabrics, folds and late sinistral and dextral shear zones can be correlated throughout the OMG, but are best exposed along the shores of Lough Cullin. The chronology present around Lough Cullin indicates that a similar sequence of deformation events occurred in all regions of the pluton. This does not infer that the same events occurred at the same time in all areas, but rather that each region underwent a similar evolution of structural style. Figure 4.23 provides a summary of the structural relations and suggests how they might be geometrically related to the shear zone which was active during their formation. The structures all provide evidence for sinistral transpression and it is suggested that they were formed in the progressive evolution of the same sinistral transpressive regime in which the OMG was emplaced.

The realization that plutons are often emplaced in a shear zone is a relatively recent idea in granitic structural studies. This is shown by the few
Fig. 4.23

Geometry of post-intrusive structures in the OMG

contractional sinistral imbricates

dextral shear bands

late dextral shear zones

late sinistral mylonites

main foliation (S)

sinistral S-C fabrics

sinistral shear bands (C)

discrete sinistral shears

contractional sinistral imbricates

late sinistral mylonites

folding of pegmatites and quartz veins
examples in the literature where the deformation of granites is geometrically described in terms of the external shear zone. The work of Davies (1980), Castro (1986) and Guineberteau et al. (1987) deals with plutons emplaced in response to shear zone deformation. Their work concentrated more on pluton scale emplacement models, with a brief description in some cases of the internal foliations which were produced in response to the shear zone deformation. There has been no attempt to establish a structural history inside each pluton and relate this to the progressive evolution of the shear zone and the changing rheology of the magma as it cooled. It can be argued that a detailed knowledge of the internal structures of the OMG provides a better understanding of the overall tectonic environment, e.g. the chronology described above provides five additional indicators that a transpressive sinistral shear regime was in operation soon after the intrusion of the OMG. If this evidence is combined with the structural evidence from emplacement structures a detailed knowledge of the shear zone deformation can be established. If, as in the OMG the shear zone deformation is classified as one major event then the granitic structures can provide information on the progressive evolution of that structure.

The model of Hutton (1982) for the intrusion of the Main Donegal Granite remains one of the few published examples of plutons in which the internal structures and, in particular, the strain history has been established. This model was based on many years of research on the Dalradian country rocks of NW Donegal and the large amount of field data produced by Pitcher and his co-workers (Pitcher & Berger 1972 for details) This study was also fortunate in making use of detailed work of Jones (1989) which produced a structural model for the Ox Mountains Inlier. The time consuming nature of the detailed fieldwork necessary may partly explain why these types of studies are not more common.
CHAPTER 5

THE L. TALT AND EASKY L. ADAMELLITES

5.1 Introduction

The purpose of this chapter is to present field data and a petrological description of the two satellite plutons which together with OMG comprise the OMIC. The Lough Talt Adamellite (LTA) and the Easkey Lough Adamellite (ELA) are emplaced in the central part of the Ox Mountains Inlier (Fig 5.1). These plutons were originally thought to be synchronous in crystallization age with the OMG, which has been dated at c480 Ma (Pankhurst et al. 1976 and Andrews et al. 1978). The work of Yardley and Long (1981) suggested that these pluton may be much younger. Long et al. (1984) showed, by Rb/Sr whole rock dating of samples from the ELA and LTA, that these plutons were significantly younger at 400 ± 33ma. This date agreed with several important observations which implied that the ELA and LTA were much younger (Long et al. 1984).

(1) The relative lack of deformation in the adamellites compared with the OMG.

(2) A sample of ELA which was included in the OMG samples, by Pankhurst (1976) fell significantly below the Rb/Sr isochron line.

(3) The limited thermal aureole around the OMG compared with conspicuous thermal metamorphic assemblages surrounding the adamellites. The OMG is considered to have been emplaced within metasediments which were close to the peak of regional metamorphism.

It is an aim of this study to establish a model for the emplacement and
deformation of the adamellites which, by detailed field mapping, would provide further critical evidence for their emplacement age relative to the tectonic framework and history of the Ox Mountains Inlier.

In the following sections, the LTA is described first followed by the ELA. Significant similarities or differences are discussed in the final section.

5.2 The Lough Talt Adamellite

5.2.1 Contact relations and metamorphism

The LTA is poorly exposed in the area in which it is inferred to be intruded. Indeed the only significant outcrop occurs at the extreme eastern edge of the pluton where there it is exposed in a cliff face approximately 1km S of L. Talt (G405 147). The contact between the LTA and the OMG is not exposed anywhere along its length. Pankhurst et al. (1976) and Long et al. (1984) followed Taylor's (1968) map in which he showed the LTA southern contact cross-cutting the OMG and extending for 0.5km into the metasediments to the south. No evidence was discovered to support such a feature during this study since this area is completely unexposed and the use of boulders to constrain near surface adamellite is of dubious value in an area where the terrain is a product of Pleistocene glacial erosion and transport. Figure 5.2 provides the results obtained during mapping of the well exposed area at the eastern edge of the LTA. The contact is apparently concordant with the metasediments of the Ummoon formation into which the LTA has been intruded. The contact is not exposed but exists in a 4-5m exposure gap between semi-pelite outcrops and uniform adamellite whose foliation trends parallel to the metasedimentary strike in this area. The metasediments along the 035° trending contact dip 68-74° to the NW. In contrast the foliation just across the contact dips 65-74° SE.
Fig. 5.2

- Foliation
- Sinistral shear band
- Stretching lineation

Structural map of the Lough Talt Adamellite

Lough Talt Adamellite

OMG

L. Hoe

N

0 100 200m
This steep contact is exposed in the face of a cliff at G405 135. The contact can be followed approximately half-way down the slope where it changes orientation sharply and becomes moderately to gently inclined to the SW. This is interpreted as a floor contact to the LTA, since the adamellite is directly above the metasedimentary country rocks. The contact trends 115° and is quite irregular with a vertical variation of up to 8m (Fig. 5.3). The metasediments below are broadly concordant to this contact.

The timing of the intrusion of the LTA relative to the regional deformation chronology was first established by the work of Andrews et al. (1978) who stated that the adamellite contact truncated $F_4$ folds (G406 133). They used this evidence to imply that the LTA and the adjacent OMG were both intruded after the development of the regional $F_4$ folds. The work of Long and Max (1977), Yardley and Long (1981) and Long et al. (1984) and Jones (1989) largely proved that the OMG is much older and was intruded at a late stage in the major $D_3$ event or the shear zone deformation. Detailed examination of the exposed contact during this study has confirmed the work of Andrews et al. (1978) and shown that the LTA was emplaced after the $F_4$ folding episode.

Yardley and Long (1981) noted that the LTA aureole extends into the country rocks in which thermal andalusite and sillimanite developed overprinting the regional staurolite grade.

5.2.2 The LTA foliation

The LTA foliation data are limited to the exposures at the easternmost part of the pluton and this can only be used in a limited way to determine the emplacement mechanism and history of this pluton.

The mapped foliation pattern exhibits a sigmoidal pattern with the foliation strike rotating from 060° to 087° in the central zone (Fig. 5.2). This region provides a partial section from the SE contact inwards which is 0.8km
Fig. 5.3

approximate horizontal scale

zone of intense L fabrics

section from the LTA contact inwards
long perpendicular to strike (Fig 5.3). The foliation changes dip inwards from the contact. It dips 65-74° SE at the contact and becomes more shallow moving towards the NW. This shallow region corresponds with the area where the foliation swings E-W. The outcrops furthest from the contact are slightly steeper than in the shallow zone and dip 36-52° SE. This foliation pattern can also be related to variations in the strain data recorded by the xenoliths (Section 5.2.3).

The foliation is concordant with the NE-SW trending contact but sharply discordant with the gently inclined floor contact. This sharply discordant fabric suggests that its current orientation may be due to a superimposed tectonic strain which modified earlier foliation patterns formed during magmatic state deformation.

The nature of the foliation when examined in both hand specimen and thin section confirms that it is the result of a solid state type deformation. This foliation is an alignment of alkali feldspar phenocrysts, biotite and quartz (Plate 5.1). Its intensity varies with the amount of shear strain present (section 5.2.3). In low strain samples the foliation is barely visible and is a weak alignment of alkali feldspar phenocrysts. The most ductile component, quartz, becomes more deformed with increasing strain and elongate ribbons of quartz become readily visible. Biotite is variably altered to chlorite and is generally elongate parallel to the foliation in more highly strained areas. In less deformed areas it has a more random orientation.

A characteristic feature of the LTA foliation is the presence of minute fractures in the feldspar phenocrysts. This is visible in hand specimen but is particularly clear in thin section where the fractures are frequently infilled with small quartz veins. The offsets are rarely greater than 1mm however the microstructural implications of their presence are important and are discussed in detail in Chapter 6. Briefly however, the brittle behaviour of feldspar
Plate 5.1.

A hand specimen of Lough Talt Adamellite which exhibits a solid state foliation formed by a shape preferred orientation of alkali feldspar, plagioclase, biotite and quartz. The pinkish/red feldspar colouration is typical of the adamellites.
Fig. 5.4

(a)

X/Z shape ratios of xenoliths in the LTA

(b)

X/Z shape ratio

Strain profile from LTA contact inwards
Plate 5.2.

The Lough Talt Adamellite is in contact with psammites of the Ummoon formation and this is likely to be a floor contact for this pluton. The adamellite displays an intense L type foliation which has formed as result of the solid state deformation being at a high angle to an earlier fabric. Lough Talt (G403 133).
provides good evidence that the adamellites were deformed at much higher crustal levels compared with the OMG. The fractures form in synthetic and antithetic systems relative to the overall shear regime discussed below.

5.2.3 Strain analysis

The LTA is the only body in the OMIC which contains significant numbers of xenoliths which can be used for strain analysis. These xenoliths are almost entirely microdioritic in composition and are therefore likely to have been originally spherical, thus they act as good indicators of relative strain throughout the pluton. The map in Figure 5.4a displays the results obtained plotted with a summary of the foliation data. A strain profile is plotted in Figure 5.4b and extrapolated towards the contact. The results are in broad agreement with the strain intensity estimated by visual observation of the degree of preferred orientation at each locality. The contact zone does not contain any xenolitic material but the degree of preferred orientation is very intense.

The higher strain zones correspond with the steeper NE striking foliation regions and the low strain values can be identified as the region with shallow E-W striking foliation. Where possible a K value has been determined from the xenolith shapes in three dimensions. These values are predominantly 1 > K > 0, which indicates that the finite strain ellipsoid has a predominantly flattened, oblate shape. In two examples the K value is > 1 and this agrees with the observation of foliations above the floor contact zone which have an intense L-type character (Plate 5.2).

5.2.4 The foliation model

The formation of linear fabrics in an isolated zone where most other foliations are planar is likely to be caused by a localized superimposition of two strain ellipsoids at a high angle to each other. The overall tectonic foliation dip is moderately towards the SE along the eastern contact this intensifies an
Fig. 5.5

Zone of intense S type foliations where solid state deformation is sub-parallel to earlier magmatic state fabric.

Solid state foliation at a high angle to the gently inclined xenoliths.

Magmatic state foliation decreases intensity inwards.

L fabric where solid state and magmatic state foliations intersect at high angles.

Approximate horizontal scale 0 100m

A model for the foliation development in the LTA.
originally steep magmatic foliation, which was probably sub-parallel to the contact. In the region above the floor contact the moderate to steeply dipping foliation may have modified an original gently inclined concordant magmatic foliation to produce an L-type foliation. Further evidence for this type of strain superposition comes from the observation of the xenoliths in the low strain E-W, striking central part of the pluton. Here many of the xenoliths are flat lying compared with a more steeply inclined weak foliation. Figure 5.5 illustrates a model for the possible production of the foliation patterns observed in the LTA. An original contact-parallel magmatic state foliation is strongly developed at the margin and quickly becomes less intense towards the centre of the pluton. This foliation is then modified by sinistral shear to produce the present foliation pattern with linear fabrics along the floor contacts and intense planar fabrics along the eastern contact zone. The more intense internal zones are due to variations in the intensity of the solid state deformation. This model has important implications for the tectonic evolution of the OMIC and in particular the sense of shear active during the deformation.

5.2.5 Shear sense indicators

Shear sense indicators are developed on various scales. The strike-swing in the central part of the exposed region and the shallow dip consistent with a model in which zones of sinistral shear modify earlier magmatic foliations. At outcrop scale shear sense criteria include S-C fabrics and the development of antithetic and synthetic fracture zones in feldspar phenocrysts.

Sinistral shear bands, or S-C fabrics are formed in the following areas:

1. In outcrops along the south-eastern contact.
2. In the L-type fabric zones above the floor contact.
3. In internal high strain zones within the LTA pluton.

The main stretching lineation associated with these S-C fabrics is a gently SW plunging elongation of quartz; all of which implies that the main foliation
is the product of sinistral transcurrent movements. A summary of the shear sense data is also presented in Figure 5.4.

There are narrow zones in the pluton in which dextral shear fabrics have formed. These deform the sinistral shear bands described above and have a more brittle appearance, which may imply that they formed at a later stage. The maximum width of these zones is 2m and their orientation is approximately parallel to the main sinistral fabric, so a geometric relationship between these zones in terms of an antithetic and synthetic relationship is ruled out. These zones are inferred to have formed after the main sinistral event which deformed the LTA.

The observations made in this study suggest that an early sinistral shear couple deformed the LTA, distorting possible magmatic state foliations close to the contacts and this was modified by later dextral shear. An emplacement model for the LTA is difficult to constrain due to the lack of exposure throughout the pluton, however it is possible to make some suggestions with knowledge of the large scale structure of the Ox Mountains Inlier. This is further discussed in chapter 7.

5.3 The Easkey Lough Adamellite

5.3.1 Contact relations and metamorphism

The Easkey Lough Adamellite (ELA) is an elliptical pluton which is orientated NNE-SSW and is approximately 5.6km long and 2.3 km wide. It is situated 10km NE of Lough Talt and is entirely intruded within the amphibolite grade Ox Mountains metasediments (Fig 5.1). This pluton is not well exposed and is mainly obscured by drift. In contrast with the LTA, isolated exposures occur throughout the pluton enabling more of a map to be made of the foliation intensities and patterns. The ELA is a uniform coarse grained
adamellite with no internal contacts. It is possible to observe the external contact relations at three localities.

(1) The contact is exposed in a stream section 400m ESE of White Rock (G467 253). Here the contact trends 023° and is a single contact which separates hornfelsed pelitic metasediments and fine to medium grained white weathering granite. The granite or adamellite contains elongate quartz grains which are aligned parallel to the strike of the contact but dip 80° SE in contrast to the metasediments which dip 40° NW.

(2) The contact is not exposed but can be mapped in a 50m exposure gap in a stream section (G461 220). The adamellite inside the contact is coarse grained with a barely discernable shallow E dipping foliation. The metasediments dip moderately to steeply to the NW and have a slightly hornfelsed appearance.

(3) The contact is not exposed but can be mapped in a 5m gully in the southern part of the pluton, 400m NW of Lough Callow (G444 202). The contact here trends 077° and separates semi-pelites and psammites from medium grained adamellite. The foliation in the adamellites is, again, strike parallel to the main cleavage in the metasediments but dips moderately SE in contrast with that in the metasediments which dips moderately NW. The foliation and cleavages are both sharply discordant with the contact at this point.

The timing of the intrusion of the ELA relative to the deformational history is considered by most previous workers to be identical to the LTA (Andrews et al. 1978, Yardley & Long 1981 and Long et al. 1984). The intrusion of granite dykes or sheets which cross-cut F₄ folds just to the S of the pluton (Andrews et al. 1978) provide evidence that the pluton was emplaced post D₄ and examination of these exposures during this study confirms and agrees with that interpretation (Plate 5.3).

A considerable metamorphic aureole exists in the surrounding metasedi-
ments and although this aspect has not been examined in detail during this study it is useful to summarize the work of Yardley and Long (1981). They found that the contact aureole of the ELA is developed in pelitic rocks which contain the following regional assemblages.

Staurolite±kyanite+garnet+biotite+muscovite±chlorite+plagioclase+quartz+rutile or ilmenite.

These may have undergone partial retrogression with the production of epidote, chlorite, chloritoid, albite and white micas. They considered this retrogression to have predated the intrusion of the ELA.

The contact assemblages according to Yardley and Long (1981) are:
(1) Andalusite+biotite+muscovite+staurolite+garnet+ilmenite.
(2) Andalusite+biotite+muscovite+staurolite+garnet+magnetite.
(3) Sillimanite+andalusite+K-spar+biotite+muscovite+ilmenite.

From these assemblages they produced a pressure/temperature estimate of conditions of the inner aureole of $595 \pm 30^\circ C$ and $2.5 \pm 0.5$ kbars, which corresponds to an emplacement depth of approximately 8km. This is in sharp contrast with the OMG which is considered to have been emplaced at 6-7 kbars depth (Yardley et al. 1979).

5.3.2 Foliation development in the ELA

The ELA foliation displays variations in its orientation and degree of development (Fig 5.6). The main part of the exposure exists in a 2km zone which extends northwards from the eastern shore of Easkey Lough (G444 231). In the southern part of this exposure the NNE striking foliation is a weak alignment of quartz and alkali feldspar which dips moderately to the ESE. This gradually increases in intensity towards the N. The northern area of these exposures at Trasgarve (G451 246) contains the most deformed rocks in the ELA and these display shear band development (see below). In this area the foliation dips steeply SE.
Structural map of the Easkey Lough Adamellite

- Contact
- Foliation
- Sinistral shear band
- Stretching lineation

Fig. 5.6

Trasgarve region displaying sinistral shear bands

Late dextral shear zone

White Rock

L. EASKEY

Lough Callow
A series of exposures 2km NE at White Rock contain a moderately developed NNE striking foliation which dips steeply ESE. This foliation cuts across the inferred trace of the contact in this region. The exposures discussed in (2) above, display a weak foliation which is gently inclined with variable strikes indicating that the levels of strain are very low at the eastern contact. The exposures in the region 400m NW of Lough Callow (G444 202) contain a moderately developed foliation which strikes NNE and dips moderately to the ESE. In a similar manner to the contact and foliation relationship in the northern region this foliation cuts directly across the contact.

In contrast with the LTA, the ELA does not contain xenoliths of any kind which means that an estimate of relative strain throughout the pluton is not readily available. The degree of preferred orientation however, can be estimated which provides a reliable indication of the areas of high strain in the pluton.

The foliation can be classified as a solid state type deformation as all components are deformed. In low strain samples quartz is the only component which is deformed and this forms slightly ovoid blebs whose shape ratio depends on the finite shear strain which each sample has undergone. Biotite and chlorite are relatively undeformed in low strain samples but highly elongate in high strain areas. Feldspar phenocrysts are aligned parallel to the foliation in low strain samples and in high strain regions displays fracture systems which are related to the principal strain orientations (Plate 5.4). In a similar way to the LTA foliation the feldspar behaviour during the deformation can provide an additional constraint on the PT conditions at the time of the foliation development (see Chapter 6). The sharp discordance of the foliation in the N and S regions suggests that the foliation formation was not directly due to the operation of internal magma processes, but rather the product of an external tectonic stress field. No evidence was discovered in the ELA to prove
Plate 5.3.

A large vein or sheet of adamellite cross-cuts $F_4$ folds close to the contact of the Easkey Lough Adamellite, implying a post $D_4$ emplacement age for this pluton. Lough Callow (G449 203).

Plate 5.4.

A hand specimen of the Easkey Lough Adamellite with a strong foliation developed by the shape alignment of quartz, biotite and alkali feldspar. This hand specimen also shows evidence for brittle cracking of feldspar and quartz. Lough Easkey (G451 232).
the existence of an earlier magmatic state deformation.

5.3.3 Shear sense indicators

Evidence for sinistral shear is provided by the observation of sinistral S-C fabric development along the western margin of the pluton. This occurs in rocks which are closest to the inferred trace of the adamellite contact on this margin (Fig. 5.6). These indicators support the model in which a shear zone has deformed the western margin of the ELA. There is no evidence in the adjacent metasediments for this shear zone, however there is a 400m exposure gap along which the possible trace of this shear zone may lie.

This deformed margin is the only part of the ELA in which evidence for sinistral shear has been observed. The sinistral shear bands deform an S foliation which is similar to that formed in the remainder of the pluton and it is considered likely that the S-C fabric formed as an intensification of this foliation rather than by production of two new fabric planes in this part of the pluton.

In the region of granite exposure to the east of Easkey Lough (G444 231) there are four narrow discrete shear zones in which dextral shear bands have been observed. The foliation swings into these zones and becomes heavily modified. The widest zone is 50cm and the adamellite in the shear is very strongly deformed to produce an ultra-mylonite (Plate 5.5). These zones have an average orientation of 050°; this may imply that they are antithetic shears to the main sinistral system. The intensity of the deformation in these zones however, suggests that there was a separate, later dextral event which could possibly be correlated with the late dextral shearing in the LTA.

Andrews (1984) suggests a model for the emplacement of the ELA in which the granite is emplaced in a dextral pull-apart structure. His main evidence for this argument was provided by data produced by the measurement of fractures in feldspar phenocrysts. He determined the shape ratios of quartz and feldspar
Plate 5.5.

A late dextral shear zone deforms the Easkey Lough Adamellite producing a fine grained mylonitic fabric (on the right) in contact with a more normal foliation. Lough Easkey (G451 232).
grains or aggregates and produced a map of strain intensity which is in close agreement with the estimates of the degree of preferred orientation made in this study (Fig 5.7).

In this work Andrews (1984) counted the total of dextral and sinistral fractures of feldspars in thin sections from three hand specimens of ELA cut parallel to the principal planes of the strain ellipsoid. His data clearly showed that dextral fractures were predominant and used this to imply a dextral shear on the whole pluton during emplacement.

His conclusions may be incorrect for the following reasons:-

(1) It is considered by this author that three hand specimens are too small a data set to make any valid inferences towards the emplacement model for the ELA.

(2) The narrow zones of dextral shear in the ELA may have formed regions of predominantly dextral fracturing in their immediate vicinities and indeed Andrew's samples were taken from the region in which these dextral shears are common.

(3) An alternative explanation for his predominance of dextral fracture is that these are 'domino' fractures which have formed in an antithetic orientation to the main sinistral shear couple. Zones of antithetic shear are common in areas of relatively rigid material (Hanmer 1986) surrounded by a more ductile material. Feldspar is much more resistant and therefore likely to fracture than the quartz and biotite matrix (Simpson 1985). Dextral shears in Figure 2d of Andrews (1984) appear to be in the correct orientation for antithetic, domino fracture relative to the main foliation.

The major conclusion from the work undertaken in this study on the ELA is in sharp contrast with the work of Andrews (1984). The main evidence suggests that a sinistral shear zone deformed the ELA at or shortly after its intrusion and this was modified by later more localized and brittle dextral
Fig. 5.7

Preferred shear direction

Modified from Andrews (1984)
deformation. The possibility that the operation of the sinistral shear zone may have influenced the site of the ELA is discussed in chapter 7.

5.4 Petrography of the adamellites

Both the LTA and ELA are composed of a very similar type of adamellite in terms of modal composition and textural relationships.

The adamellites are coarse grained except in the contact region of the ELA where they are medium- to fine-grained and white in colour. Both plutons are internally homogenous, (apart from the xenolithic material in the LTA), light grey or white weathering, with up to 4cm pinkish-red alkali feldspar megacrysts which give a characteristic appearance to the hand specimen. Plate 5.4 is of a hand specimen taken from the western part of the ELA which has a moderately strong foliation development.

The LTA and ELA are principally composed of alkali feldspar, plagioclase, quartz and biotite. The alkali feldspar content varies from 52-75% of the total feldspar content.

Plagioclase occurs as single crystals or composite aggregates which are equant, euhedral to subhedral and up to 2mm in size. The average composition is oligoclase (An20), with a faint normal zoning towards the margins of each grain. The twinning is predominantly simple, with the development of faint albite twins in most crystals. Plagioclase inclusions are biotite and chlorite, epidote and muscovite. Plagioclase in most cases has a dusty appearance mainly due to alteration to sericite.

Alkali feldspar is present as up to 4cm equant, subhedral crystals. It is composed of orthoclase which exhibits simple twinning, or is untwinned and is microperthitic in most cases. The orthoclase crystals contain inclusions of plagioclase and biotite towards their margins. Both alkali feldspar and
plagioclase display evidence of intracrystalline fracturing which may contain quartz veins.

Quartz is present as elongate aggregates in those examples in which a foliation is developed. In the undeformed state quartz crystals are up to 3mm in size and display undulatory extinction.

Biotite occurs as large (up to 3mm) aggregates which contain subhedral laths. The biotite is associated with opaques and commonly altered to chlorite.

Accessory minerals are epidote, muscovite and allanite.

The adamellite texture is largely a result of the deformation which occurred following its crystallization. Even samples which do not appear to contain a foliation show evidence of undulatory extinction in quartz and evidence for brittle fracturing in feldspar. The deformation mechanisms and microstructures which characterize the adamellite foliation is more fully discussed are chapter 6.

5.5 Summary and comparison

The LTA and ELA display a remarkable number of similarities which may imply that they underwent similar intrusion and deformation histories.

These are:-

(1) The modal compositions, petrography and textures are almost identical. The exception being the presence of xenoliths in the LTA.

(2) Samples from both plutons 'fit' on the same Rb/Sr whole rock isochron which produces an age date of c400ma (Long et al. 1984).

(3) Their structural history is quite similar; both show the operation of a sinistral shear zone during the period of main foliation development. This was most intense at the margin adjacent to the Glennawoo shear zone which is a major structural discontinuity in the Ox Mountains Inlier. They both show
evidence for later localized, more brittle, dextral deformation. A difference between the two is that evidence for the presence of a magmatic state foliation exists in the LTA and not in the ELA.

(4) In both adamellites similar deformation mechanisms were operative during the foliation production event (see chapter 6).

(5) Both plutons are situated in a releasing bend site relative to reactivated sinistral movements on the Glennawoo shear zone (see Chapter 7).
CHAPTER 6

SUB-SOLIDUS TEXTURAL DEVELOPMENT IN THE OMIC

6.1 Introduction

The purpose of this chapter is to present a detailed description and analysis of the textures and microstructures which dominate the OMIC fabrics. This is based on hand specimens and thin sections taken from the OMG and the adamellites. The purpose of this description is to attempt to recognise the deformation mechanisms which were operative in each mineral phase during the deformation of the OMG and the adamellites. Recognition of the deformation mechanisms which are operating during a deformation process can provide information on the characteristics of the flow regime in which that process is taking place. These characteristics depend on a variety of environmental factors (Knipe 1989); temperature, pressure, shear stress and fluid pressure in particular and also lithological controls such as mineralogy, porosity and permeability. Figure 6.1 is taken from Knipe (1989 Fig. 1) which provides a good summary of the inter-relationships between lithological, environmental controls and material processes during deformation.

It is of particular interest in this study to attempt to ascertain the possibility that the recognition of deformation mechanisms could lead to an estimation of the pressure and temperature conditions prevalent during deformation of the OMG. The work of Simpson (1985) suggested that crystal plastic deformation mechanisms in feldspars are characteristic of amphibolite facies deformation and that fracture or cataclastic flow processes in feldspars are typical of the greenschist facies. The emplacement depth and
Fig. 6.1

LITHOLOGICAL CONTROLS:
mineralogy, porosity, permeability, etc.

ENVIRONMENTAL CONTROLS:
temperature, pressure, shear stress, fluid pressure, etc.

SELECTION OF DOMINANT MATERIAL PROCESSES.

DEFORMATION MECHANISMS

MATERIAL PROCESSES
Within grains:
1. Diffusion
2. Dislocation movement
3. Twinning
4. Elastic distortion
5. Fracture
6. Grain boundaries
7. Reaction
8. Diffusion
9. Crystal growth
10. Fracture
11. Frictional sliding
12. Sliding by diffusion or dislocation processes

from Knipe (1989)
PT conditions are not fully constrained for the OMG due to the poor development of a thermal aureole and therefore the recognition of the deformation mechanisms may provide additional constraints on emplacement depths and the environmental conditions prevalent during the sub-solidus deformation of the OMG.

The review of deformation mechanisms by Knipe (1989) divided them into three main categories.

(1) Diffusive mass transfer

These processes include those in which strain is accumulated by redistribution of material by diffusion during deformation. The diffusion mechanism usually takes place from zones of high intergranular normal stress to areas of low normal stress and can take place via a number of pathways:

(i) Through the crystal (Nabarro-Herring creep).

(ii) Along distorted and disordered solid-solid grain boundaries (Coble creep).

(iii) In a thin film along grain boundaries (pressure solution).

(iv) In a bulk fluid which may itself be experiencing flow.

(2) Crystal plasticity

This involves the accumulation of strain by intra-crystalline processes such as the movement of dislocations and the process of twinning. Dislocations are crystal lattice defects which are controlled by the crystal structure and by impurities. At low temperatures, (< 0.5 of the melting temperature in laboratory experiments), deformation takes place by dislocation glide where dislocation motion is confined to slip planes. This leads to dislocation tangles and 'work hardening' which is usually closely followed by fracture. At higher temperatures (> 0.5 melting temperature) thermally activated accommodation processes such as dislocation climb, which is the movement of dislocations out
of slip planes, reduce the work hardening effect. The process of dislocation climb may occur by either dynamic recovery or dynamic recrystallization and these processes have been described by White (1976) and Tullis & Yund (1985) and are summarized briefly below.

Dynamic recovery produces deformation lamellae and sub grains by the ability of the dislocations to climb onto adjacent glide planes. In sub-grain production, grains of any initial size deform homogeneously in steady state flow and the dislocations so formed are uniformly distributed into low energy walls to form the sub-grains. In materials undergoing recrystallization accommodated dislocation creep, climb is difficult and the dislocations become pinned. Grain boundaries migrate into areas of higher strain within the lattice in response to dislocation gradients. Bulges in the grain boundaries may pinch off to become strain free grains. Dynamic recovery processes are considered to require higher temperatures (Tullis & Yund 1985) and may also produce recrystallization by the continued misorientation of sub-grain boundaries to > 10°, thus changing low angle sub-grain boundaries into high angle mobile grain boundaries.

The formation of strain free new grains may be followed by the development of new dislocations to accumulate further strain, which results in dislocation climb and thereby new sub-grains by further dynamic recovery. Knipe (1989) noted that this cyclic process often controls the development of shear zones by the localization of deformation and the reduction of grain-size. This can lead to the dominance of diffusive mass transfer processes, which are more grain size dependant than crystal plastic mechanisms.

(3) Frictional sliding, fracture processes and cataclasis.

Knipe (1989) distinguishes frictional sliding processes where fracture does not dominate, from true fracture processes.

(i) Frictional sliding
This occurs where grains slide past each other without internal deformation. This process is very pressure dependant and operates in regimes of low confining pressures or high fluid pressures, and is therefore common in partially or unlithified sediments, or in fault zones containing incohesive gouge.

(ii) Fracture processes and cataclasis

These involve the nucleation, propagation and displacement along new surfaces created during deformation. The material may fragment, rotate and dilate or deform by grain boundary sliding in the process of cataclasis which dominates faulting at high crustal levels.

The operation of deformation mechanisms in any of the above categories can be predicted by the observation of microstructural features in the OMG and the adamellites and this is presented in the following sections.

The other major textural change which may take place accompanying microstructural development is the production of new phases in response to deformation. This is essentially a metamorphic reaction process. In the OMG the growth of new phases can be geometrically related to the stress field which produced the fabric. These features are described along with a discussion of their implications for the deformation and rheological changes in granitoid rocks.

6.2 The microstructural features of the discrete sinistral shears

The discrete sinistral shears described in section 4.2 are considered to be the first structures which formed in the OMG in the transition from the late crystallization to the early sub-solidus deformation. Thin sections from these zones were examined to obtain microstructural information and to determine whether the microstructure survived the later deformation which produced the main foliation.
The hand specimens taken from these zones are fine to medium grained with a moderate to strong planar alignment of biotite and feldspar. These narrow shear zones have been very strongly deformed, and must contain extremely high shear strain values.

After initial inspection in thin section, these rocks do not appear to be highly deformed, having the appearance of a moderately foliated microgranite. They do however, show certain features which may be a product of the high shear strains.

The microstructure in plagioclase in the discrete sinistral shears is unique within this pluton. Plagioclase displays widespread sub-grain development and is deformed to elongate ribbons or aggregates (Plate 6.1). These ribbons are elongate parallel to the foliation and adjacent quartz ribbons but the individual grains do not show any internal deformation features, which would suggest that they were strongly deformed. The grains show albite twinning and alteration to sericite, but do not show zoning, which suggests that they have been recrystallized.

Quartz is completely recovered to form elongate ribbons or aggregates which in general do not show internal deformation, however some grains do show a slight undulose extinction, perhaps related to later deformation.

Biotite is present as elongate aggregates of recrystallized laths which are associated with blebs of opaques which may have exsolved from biotite during recrystalization.

The thin sections taken from the discrete sinistral shear zones all display an annealed, medium-grained texture which suggests that deformation took place under high temperatures and pressures. There are two alternative explanations for the formation of the discrete sinistral shear zones.

(1) Fracturing or cataclasis of partially solid magma. A crystalline framework may exist which is capable of deforming in a crystal plastic manner. The
Plate 6.1.

Elongate plagioclase and quartz ribbons in a thin section taken from a discrete sinistral shear which suggests the operation of high temperature shearing. The plagioclase crystals are un-zoned which indicate that they have been recrystallized and annealed. Field of view 4mm. Pontoon (G216 049).
inter-crystalline melt may promote fracturing if the pore pressure is increased by tectonic 'squeezing' to overcome $\sigma_3$. Experimental deformation of partially crystalline granitic aggregates by Dell'Angelo and Tullis (1988) suggests that samples with 15% melt will undergo cataclasis because the excess melt cannot flow fast enough laterally to accommodate stress. In this situation the crystalline framework will fracture and be translated on a film of melt which may later crystallize to form a fine grained microgranite along the shear plane.

(2) The shears are narrow zones in which superplastic deformation mechanisms such as grain boundary sliding are operative in a consolidated, high temperature, granitoid which is deforming by crystal plastic processes. This implies that the fine grained product of the crystal plastic deformation process may behave in a superplastic manner. Behrmann & Mainprice (1987) suggest that superplastic behaviour is likely to dominate if the grain size is locally reduced to 10 microns. In the OMG, such grain size reduction resulting from crystal plastic deformation has been observed in the development of the S-C fabrics (see 6.3) as a localization of the main foliation.

Ductile 'steady state' flow will occur if the crystal lattice is deforming by recovery accommodated dislocation creep. This is the expected process during high temperature deformation of the granitoid. The development of discrete displacements in a ductile regime has been discussed by Hobbs et al. (1986). They suggested an alternative mechanism to localized melting phenomena for the development of pseudotachylytes in quartzites, and noted that it may apply to other rock types as well. Hobbs et al. (1986) defined a critical temperature ($T_c$) at which transient work hardening equals the product of thermal fluctuation and the heat generated by shearing. A ductile instability will form if the temperature is below $T_c$; then the rate of change of stress with respect to strain is negative. The average shear strain in the zone is concentrated in the instability, creating zones of very localized high shear
strain which are then overprinted by further ductile shearing. Hobbs et al. suggest $T_c$ for a diabase is 250° C (?). The OMG shears however, are thought to have formed close to the granodiorite solidus which is approximately 650° C at 6Kbars (Wyllie 1977) nevertheless a similar process at higher temperatures may account for the formation of the discrete shears in the OMG.

There is no evidence preserved to suggest that the discrete sinistral shears formed as a localization of the main ductile pervasive foliation because it clearly overprints the shears in all examples. This implies that the discrete shears formed by either fracturing of semi-crystalline magma or a high temperature ductile instability.

6.3 Microstructural characteristics of the main foliation in the OMG

In the following sub-sections microstructural features of each of the major components of the OMG and adamelites are described in turn. These are; feldspar, quartz, biotite/chlorite and amphibole.

6.3.1 Feldspar microstructure in the OMG

Plagioclase and alkali feldspar exhibit differing microstructures which indicate that in the prevailing environmental conditions they had a different rheological response to the external stress field which existed during the formation of the main foliation.

Whereas plagioclase is generally more equant, alkali feldspar is elongate in all hand specimens and thin sections. This implies that plagioclase behaved in a more resistant manner than alkali feldspar. Evidence for crystal plastic deformation of plagioclase however, is provided by the observation of large sub-grains within the crystals. These divide some of the crystals up into several large sectors which exhibit different extinction positions during
rotation of the microscope stage (Plate 6.2). These sectors may have formed by slight rotation of individual sub-grains which were bounded by conjugate kinks in the internal lattice structure of the plagioclase crystals and are not believed to be 'true' sub-grains formed by dislocation creep (Tullis & Yund 1983). Further evidence of the rigid behaviour of plagioclase is provided by the presence of antithetic dextral shear zones which split originally single crystals into two (Plate 6.3) The original fracture may have formed in response to high normal stresses or high strain rate which the plagioclase grain could not accommodate by crystal plastic deformation. The fracture was initiated in an antithetic orientation and further dextral shearing rotates the plagioclase crystal by a domino fault mechanism (Fig. 6.2). These features are not widely observed and in most cases plagioclase behaves in a passive manner and other components accommodate the stress.

In thin sections taken from the higher strain, mylonitic lithologies there is a suggestion that resistant plagioclase may rotate or roll in a ductile quartz flow regime (Plate 6.4). This process can explain the existence of sub-rounded plagioclase crystals which, at the cessation of the main shearing event, have their long axis perpendicular or at a high angle to the foliation. Plate 6.4 illustrates this process and shows that the perturbation effect which the plagioclase has on the surrounding quartz/feldspar mylonite will be most highly developed when the plagioclase long axis is perpendicular to the foliation.

Plagioclase may also develop recrystallized sub-grains at the margins of crystals which are otherwise internally undeformed. This may be caused by the build up of dislocations at the margins of plagioclase crystals in response to high normal stresses followed by a recrystallization either by sub grain rotation or by grain boundary migration. The production of a finer grain feldspar mylonite may facilitate a switch in deformation mechanisms from crystal plastic creep to one in which strain is accumulated by grain boundary
Antithetic fracture in plagioclase
in response to high normal stress
Plate 6.2.

Large sub-grains developed in plagioclase which may be bounded by large kinks in the crystal lattice. This is evidence for crystal plastic deformation in plagioclase during the main foliation development. Field of view 4mm.

Plate 6.3.

A plagioclase crystal which has fractured and been displaced with a dextral sense of shear which is antithetic to the overall shear sense. Field of view 4mm.
sliding (White et al. 1980).

In summary plagioclase behaviour during the main foliation development is characterized mainly by crystal plastic deformation, either internally by kink formation, or at the margins by a dynamic recrystallization process. In high strain zones the plagioclase may behave as rigid particles rolling in a quartz mylonite, or they may fracture if the normal stress becomes too large.

Alkali feldspar exhibits considerably more evidence of its response to the external stress field. The main feature is that it is much more elongate than plagioclase. Plate 6.5 illustrates an alkali feldspar crystal from a medium to high strain mylonite from the Pontoon region (G205 071). There are no visible fractures or offsets of crystal margins which implies that deformation took place by a crystal plastic mechanism or a combination of both processes.

Internal deformation features are common in alkali feldspar. The main feature is an uneven extinction which has a different appearance depending on the triclinicity of the alkali feldspar. Microcline may show simple uneven extinction as in Plate 6.6, or more complex examples in which the characteristic microcline twinning is heavily deformed. Not all alkali feldspar is inverted to microcline and examples in which partial inversion has occurred have been observed in thin sections of granodiorite. Bell and Johnson (1989) suggest that alkali feldspar triclinicity increases with increasing strain because the movement of dislocations allows increased Si/Al interchange rates and increased water access. It was not possible from this study to establish a relationship between microcline inversion and the degree of strain. Orthoclase also shows uneven extinction and sub-grain formation close to its margins which indicates that it was deforming by a crystal plastic deformation mechanism which was either dislocation glide or dislocation creep. The latter mechanism is suggested from the lack of evidence for fracturing processes and the generally elongate shape ratios, which indicate that work hardening did not occur.
Plate 6.4.

Relict plagioclase crystals in a mylonitic ground mass of quartz and feldspar. Here, plagioclase is likely to be behaving in a passive manner and rotating in the matrix. Rotation of elongate crystals perturbs the overall flow regime creating open folding of the mylonitic fabric. Field of view 4mm.

Plate 6.5.

A microcline crystal which is elongate parallel to the main foliation and also exhibits an uneven extinction suggesting that the crystal lattice is strained. Myrmekite is also developed on the margin at right angles to adjacent to the maximum principle stress direction. Field of view 4mm.
The other major textural feature which is associated with alkali feldspar deformation is the formation of fine grained plagioclase at grain margins. These plagioclase grains are characteristic of the microstructure of the shear bands which form in the widely developed shear band cleavages. The plagioclase ribbons are parallel to the quartz ribbons, but are generally more fine grained (Plate 6.4). Plates 6.5, 6.6 and 6.7 illustrate these features. Plate 6.5 displays the geometrical relationship between the foliation and the development of the small plagioclase grains. The plagioclase is concentrated in the region of the margin of the alkali feldspar which undergoes the maximum normal stress. The zones at the elongate tips of the alkali feldspars are in area of relatively low normal stress and do not exhibit these features. Plate 6.5 illustrates the fine grained nature of plagioclase aggregates compared to quartz.

Alkali feldspar replacement by plagioclase may take place by a process of myrmekite production. This was first described by Simpson (1985) who related the myrmekite geometrically to the shear zone which deformed a granite and suggested that it was characteristic of deformation of granitic rocks in the lower amphibolite facies. Plate 6.7 illustrates an orthoclase crystal with an obvious shear band which has cut the crystal in two parts. The shear band has a zonal arrangement with a central recrystallized plagioclase and quartz mylonite. On either side of this thin zone are two strips of myrmekite pools. This represents the initial stage in which shear deformation has induced myrmekite production and this is further deformed in the central zone causing mobilization of quartz and production of a fine grained recrystallized texture. This process of myrmekite production by deformation is evidence for the operation of diffusive mass transfer processes at sites of high normal stress in the foliation (Simpson & Wintsch 1989). Volume loss in these positions leads to a shape ratio increase, or flattening of the alkali feldspar, parallel to the
Plate 6.6.

An orthoclase crystal which displays uneven extinction and small plagioclase sub-grains. Field of view 4mm.

Plate 6.7.

An orthoclase crystal which has been divided by a shear zone. The shear zone displays a zonal arrangement in which myrmekite pools are formed adjacent to the orthoclase. This is then further deformed in the shear zone to form small recrystallized plagioclase and quartz grains. Field of view 4mm.
foliation. It is however uncertain if a corresponding elongation occurs at the sites of low normal stress. This process is thought to contribute significantly to softening mechanisms which produce localization of strain into shear band or mylonite zones (see section 6.4).

6.3.2 Feldspar microstructure in the ELA and LTA

Feldspar microstructure in the adamellites are dominated by fracture and cataclastic textures. In thin sections taken from the lower strain variants in the LTA and ELA feldspar exhibits little internal deformation. The exception to this is that some plagioclases show brittle disruption of albite twinning (Plate 6.8). At higher strains this disruption is dramatically increased and a cataclastic texture containing feldspar aggregates in a quartz and mica matrix is formed (Plate 6.9). No evidence was observed to suggest that either plagioclase or alkali feldspar display microstructure which was formed by crystal plastic deformation mechanisms. Both display fracturing and cataclasis in sharp contrast with feldspar behaviour in the OMG.

6.3.3 Quartz microstructure in the OMG

Quartz is generally accepted to be the most ductile mineral in quartzofeldspathic rocks at most temperatures and pressures encountered in the crust (Tullis 1985). In the OMG, LTA and ELA quartz is deformed in a crystal plastic manner which produces elongate ribbons in high strain zones. Information on the environmental factors which governed the deformation mechanisms and therefore the microstructure produced is available by observation of the operation of recovery processes to form recrystallized grains. Quartz in the OMG forms very elongate ribbons or 'stringers' of fine grained (0.1-0.25mm) aggregates (Plate 6.10). Individual grains have a generally even extinction indicating that they contain low dislocation densities. Some do however show, a slight undulose extinction indicating that the crystal lattice is slightly strained by late deformation.
Plate 6.8.

A plagioclase crystal in the Easkey Lough Adamellite which exhibits a brittle disruption of the albite twinning. Field of view 4mm.

Plate 6.9.

A highly strained example from the Easkey Lough Adamellite which displays a cataclastic texture developed in orthoclase. Field of view 4mm.
In detail the OMG granodiorites, granites and tonalites exhibit varying degrees of recovery in quartz. Thin sections taken from the NE end of the pluton from the muscovite granite at Knocknasliggaun (G372 152) show microstructural features which indicate that they underwent less dynamic recovery than the granodiorites to the SW. The features displayed in Plate 6.11 are:-

(1) Undulose extinction indicates that the quartz lattice is strained.
(2) Sub-grains, newly recrystallized grains and a transition are preserved.
(3) Grain boundaries display bulges indicating that the grain boundary migration process was non-uniform.

The formation of sub-grains and undulose extinction indicate the operation of dynamic recovery accommodated dislocation creep and the bulges on the grain boundaries indicate that some dynamic recrystallization was also operative although not dominant.

Most thin sections from the remainder of the OMG do not show the above features but instead exhibit the following characteristic microstructures in quartz.

(1) Strain free new grains with even extinction.
(2) Grain boundaries which are straight
(3) Grain boundary triple points which approach the equilibrium 120° angle (Plate 6.12).

This microstructure indicates that dynamic recovery processes were capable of relieving the hardening process and accumulated strain by steady state flow. The sub-grains become sufficiently mis-orientated to become new grains. Grain boundaries migrate in a uniform manner and may evolve towards textural equilibrium by a secondary recrystallization following deformation. This reduces grain boundary energy by growth of those boundaries with the greatest mis-orientation. Dislocation creep processes associated with recovery processes are normally favoured by higher thermal conditions and this indicates
Plate 6.10.

Quartz ribbons formed in the Ox Mountains granodiorite. Field of view 4mm.

Plate 6.11.

Quartz microstructure in the Ox Mountains Granodiorite at Knocknasliggaun (G372 152). Here the quartz displays uneven extinction and the formation of sub-grains. Field of view 4mm.
that the SW part of the pluton was probably deformed at a deeper level, or closer to the granite solidus, than the NE.

The geometry of grain boundary bulges in rock analogues has been examined by Drury and Humphreys (1988). They were able to demonstrate that these bulges had an asymmetry which reflected the local sense of shear. This occurred by a combination of grain boundary migration and grain boundary sliding (Fig 6.3). The grain boundary bulges in thin sections from the muscovite granite in the NE OMG (Plate 6.13) do display some asymmetry indicating that this criteria may be applicable in real granites where it may be useful in the absence of other shear sense criteria and particularly where undulose extinction makes quartz c-axis determination difficult.

### 6.3.4 Quartz microstructure in the ELA and LTA

The quartz microstructure in thin sections taken from the two adamellites displays evidence for the operation of crystal plastic deformation mechanisms. There was no detectable evidence for fracture processes in any thin sections taken from the normal foliation in either pluton. Quartz however, in the late dextral semi-brittle shear zones displays evidence for fracturing in the form of crystal margin offsets and the formation of linear bubble trails transecting quartz grains.

In low strain examples the quartz microstructure consists of large, elongate (up to 0.5mm long) sub-grains and deformation lamellae. The grain boundaries are serrated on a very fine scale and contain minute (0.01mm) sub-grains (Plate 6.14).

Higher strain varieties exhibit an extreme elongation of quartz with fine grained 0.01mm sub-grains occurring in the mantles which surround quartz cores. The cores display deformation lamellae (Plate 6.15). These microstructural features indicate that quartz deformation in the ELA and LTA took place predominantly by recrystallization accommodated dislocation creep and
Figure 6.3.

Mechanism of formation of grain boundary bulges. (a) Deformation at the grain boundary is accompanied by sliding and non uniform strain adjacent to irregularities. (b) Local grain boundary migration consumes zones of high defect density producing asymmetric bulges. (c) Further deformation along the grain boundary is accomodated by sliding along short segments and shear in the grain mantle. (d) Shear in the mantle modifies the shape of the bulges amplifying the bulges with the same sense of shear as the sliding. (e) Predicted geometry of grain boundary bulges for simple shear (f) low strain, (g) high strain.
modified from Drury and Humphreys (1988)
Plate 6.12.

The typical quartz microstructure in the Ox Mountains Granodiorite. This is part of a quartz ribbon and displays an even extinction and grain boundary triple points which are approximately $120^\circ$. Field of view 1mm.

Plate 6.13.

Asymmetric grain boundary bulges formed on quartz sub-grains boundaries in a thin section from Knocknasliggaun (372 152). (Field of view 1mm).

Deformation lamellae developed in quartz and small grains formed in new sub-grain boundaries. This is the typical quartz microstructure in the adamellite plutons. Field of view 4mm.

Plate 6.15.

An example of the quartz microstructure associated with the higher strain zones in the adamellites. Here the quartz sub-grains are extremely elongate and separate quartz cores which display deformation lamellae. Field of view 4mm.
recovery processes did not play a major role.

6.3.5 Biotite microstructure in the OMG

The microstructural features described for biotite in this section also include chlorite which variably replaces biotite in most thin sections. Muscovite is present in thin sections taken from the muscovite granites and is formed as a reaction product in the deformed granodiorites and granites (see section 6.5).

Biotite forms recrystallized aggregates in lower strain examples. Internal deformation features such as sub-grains or kink bands have not been observed here. Grain boundaries are straight with no bulge or sub-grain development and the aggregate triple points are low angle (Plate 6.16). This texture provides evidence that biotite deformed by a crystal plastic process accompanied by recrystallization. Thin sections from the NE portion of the pluton which show poor quartz recovery also exhibit a recrystallized biotite texture although this is not as well developed as that further SW. In these thin sections small 0.01mm sub-grains lie along grain boundaries indicating that grain boundary migration has not been complete. Plate 6.17 demonstrates typical biotite microstructure in thin section taken from the OMG with a more intense foliation development. Here biotite is recrystallized in very elongate aggregates parallel to quartz ribbons.

6.3.6 Biotite microstructure in the LTA and ELA

In the adamellites biotite displays a notably different microstructure compared with that developed in the OMG. In low strain examples open kinks deform [001] cleavage planes. Other examples show ribbon type structure in which narrow (0.2mm) strips or laths of biotite have their [001] cleavage planes inclined at 30° to the lath boundary. Adjacent laths have their cleavage planes inclined towards one another across the boundary (Plate 6.18). The boundaries are composed of fine grained (0.01mm), sub-rounded sub-grains of biotite
Plate 6.16.

This is an example of biotite microstructure in a sample of the granodiorite with relatively low strain. The biotite takes the form of aggregates of recrystallized sub-grains. Field of view 4mm.

Plate 6.17.

Elongate biotite sub-grains in a highly deformed sample of Ox Mountains Granodiorite. Biotite in this case is likely to be deforming by slip along its [001] cleavage plain. Field of view 4mm.
and ilmenite. Simpson (1985) suggested that this microstructure is analagous to that of chevron fold formation as an end-product of kinking in a multi-layered sequence. Dynamic recovery mechanisms operate in the boundaries to form sub-grains.

Some samples with higher strain foliations display folded cleavage planes and shear parallel to the [001] cleavage planes (Plate 6.19). The folded biotite indicates that recrystallization processes were not operative in biotite during deformation, in contrast with the OMG where biotite is recrystallized to straight laths.

6.3.7 Amphibole microstructure in the OMG

Widespread amphibole is present only in the diorites, however minor amounts occur in the tonalites and granodiorites. Amphibole is considered to be one of the strongest minerals and it is difficult to deform by crystal plastic processes (Nicolas & Poirer 1976). In the Ox Mountains diorites, the amphibole is deformed in shear bands which formed during the main foliation development (Plate 6.20). Individual amphibole grains are on average 0.5mm long and display a recrystallized texture with no internal features apart from a simple twinning developed in some grains. Plate 6.21 illustrates the recrystallized texture which is developed in several samples to produce aggregates of amphibole. Each grain is internally undeformed but does display a shape preferred orientation which is parallel to that of the overall aggregate and the main foliation. This microstructure is typical of that of amphibole undergoing recrystallization accommodated dislocation creep. This is considered to have developed at high temperatures close to the solidus temperature of diorite. Plagioclase in these diorites shows a similarly recrystallized texture.
Plate 6.18.

Biotite microstructure in the Easkey Lough Adamellite. Here biotite is deformed by a series of kinks. There are small sub-grains developed along these kink planes. Field of view 4mm.

Plate 6.19.

Biotite microstructure in a more highly strained sample of the Easkey Lough Adamellite. Here biotite is kinked and folded and deformed by shearing along the cleavage planes. Field of view 4mm.
Plate 6.20.

A shear band deforming hornblende in a sample of diorite from the Ox Mountains Granodiorite. Field of view 10.8mm.

Plate 6.21.

Hornblende aggregates in a sample of diorite from the Ox Mountains Granodiorite. The hornblende has undergone crystal plastic deformation in association with recrystallization to form sub-grains. Field of view 4mm.
6.4 Deformation localization and the late shear zones

6.4.1 The granodiorites and granites

The nature and geometry of the late sinistral and dextral shear zones has been discussed in section 4.6 & 4.7. These are likely to have formed by deformation localization during the late stages of the foliation development.

In hand specimen, these zones display a mylonitic fabric which is a very strong planar alignment of feldspar, biotite and quartz. There is a well developed sub-horizontal stretching lineation in all these mylonite zones. These zones characteristically are darker in appearance when compared with the surrounding zones (See Plates 4.15 & 4.16). This is predominantly caused by disseminated, fine grained phyllosilicate (mainly chlorite) which occurs throughout these mylonite zones (Plate 6.22).

Feldspar displays a fine grained microstructure with relict phenocrysts which display an asymmetric augen structure (which indicate sinistral shear) in a matrix of recrystallized feldspar, quartz and chlorite (Plate 6.22). Plagioclase often displays evidence that relict phenocrysts have been rotating in the matrix as described in section 6.3.1. Alkali feldspar on the other hand deforms by a process of sub-grain production adjacent to grain margins, related to myrmekite production. Quartz microstructure in these mylonites displays the development of fine grained, recrystallized ribbons or aggregates (Plate 6.22). Some thin sections show undulose extinction and this, combined with the fine grained nature of the quartz fabric indicates that recovery processes were not predominant over recrystallization during microstructure production.

Biotite or chlorite displays an unusual microstructure in these mylonites. In contrast with the development of elongate aggregates (as in the main foliation) it is present in disseminated form as single elongate grains or minute aggregates separating quartz or feldspar ribbons.
The microstructure described above may indicate that feldspar and quartz deformation was dominated by recrystallization accommodated dislocation creep. The operation of this process, in contrast to that of recovery, accounts for the fine grain size which may have assisted the deformation localization process by accumulation of strain due to grain boundary sliding. The switch from recovery processes in feldspar in particular may be controlled by a drop in temperature (Tullis & Yund 1985). This is further discussed in section 6.5.

6.4.2 Deformation localization and the tonalites and diorites

The tonalites, in contrast with the granodiorites, contain very low modal percentages of alkali feldspar, the diorites do not contain alkali feldspar or quartz. One implication from the differing mineralogy is that these lithologies are likely to have different rheologies during deformation because there is little alkali feldspar to induce deformation localization (see below). This begs the obvious question: does deformation localization occur to the same extent in these rocks and which are the main minerals that are causing it?

Examination of thin sections from the OMG tonalites indicates that shear bands are not as intensely developed and that they have a generally finer grain size which would cause them to deform more homogeneously than the granodiorite. This may be because the strain in the tonalites is lower since these rocks were intruded at a later stage in the deformation history. Shear bands in the tonalite mainly form by slip along [001] planes in biotite which may be adjacent to quartz ribbon (Plate 6.23) or narrow zones of fine grained recrystallized plagioclase. These recrystallized zones are much narrower than those developed in the granodiorite, averaging 2-3 grains wide, although the displacement along these shear zones may be relatively large. Small, recrystallized plagioclase feldspars are likely to have formed at the margins of the plagioclase phenocrysts and they may deform by grain boundary sliding or remain undeformed, but translated by slip along the biotite cleavage planes.
Plate 6.22.

A thin section taken from a highly strained mylonite zone in the Ox Mountains granodiorite. A medium to fine grained fabric is present, formed from ribbons of quartz and feldspar with disseminated biotite and chlorite throughout. Field of view 4mm.

Plate 6.23.

This microstructure is typical of that in the tonalites and consists of largely undeformed plagioclase crystals separated by shear bands. These shear bands are quartz ribbons and elongate biotite ribbons. Field of view 4mm.
Intensification of this deformation culminated in the development of very elongate biotite and quartz ribbons in the shear bands which separate the plagioclase phenocrysts. The phenocrysts appear to have been relatively inactive during this stage of the deformation. The localization does not appear to take place to the same extent as in the granodiorites and granites and forms more uniformly distributed shear bands rather than the intensely localized mylonite zones in the granodiorite.

It is difficult to define the environmental conditions under which this deformation took place, nevertheless the presence of a small number of sub-grains of plagioclase and small amounts of recrystallized plagioclase in some of the shear bands indicate that the deformation at least initiated under lower amphibolite facies conditions (Simpson 1985).

6.5 Deformation induced alkali feldspar replacement by myrmekite

This section deals with the characteristic microstructure which formed during the shear band formation. This is essentially a localization feature (White et al. 1980) and is further intensified to produce localized zones of mylonite which divide some of the outcrops into large blocks (see section 4.6).

The microstructural characteristics of the shear zone mylonites have been described in the previous section. There is a basic similarity between these and the shear bands in that both are largely composed of fine grained recrystallized feldspar grains which is mainly plagioclase. It has been suggested in section 6.3.1 that fine grained plagioclase aggregates form adjacent to alkali feldspar megacrysts by a process involving the production of myrmekite. This process may be related to the stress field which was prevalent during the formation of the main foliation. The formation of myrmekite in granitic rocks deformed at
lower amphibolite facies has been discussed by Simpson (1985) who suggested a model in which strain enhanced the diffusion of $\text{Na}^+$ ions towards the grain margin in a direction at right angles to the maximum normal stress.

Phillips (1980) selected two fundamental hypotheses to explain the production of myrmekite.

(1) The Becke hypothesis which interprets the quartz and plagioclase intergrowth as a product of the reaction of late stage Na and Ca bearing solutions with alkali feldspar. Albite and Anorthite are formed by:

$$\text{KAlSi}_3\text{O}_8 + \text{Na}^+ = \text{NaAlSi}_3\text{O}_8 + \text{K}^+$$

Orthoclase \hspace{1cm} Albite

$$2\text{KAlSi}_3\text{O}_8 + \text{Ca} = \text{CaAl}_2\text{Si}_2\text{O}_8 + 4\text{SiO}_2 + 2\text{K}^+$$

Orthoclase \hspace{1cm} Anorthite \hspace{1cm} Quartz

Albite and anorthite mix to give a sodic plagioclase and the quartz precipitates as vermicular quartz. This model has been extended to include the association of muscovite and myrmekite with the destruction of alkali feldspar by Phillips et al. (1972).

$$\left(\frac{3\text{KAlSi}_3\text{O}_8}{\text{NaAlSi}_3\text{O}_8}\right) + \left(\frac{\text{CaAl}_2\text{Si}_2\text{O}_8}{\text{NaAlSi}_3\text{O}_8}\right) + \text{H}_2\text{O} \rightarrow \left(\frac{\text{CaAl}_2\text{Si}_2\text{O}_8}{2\text{NaAlSi}_3\text{O}_8}\right) + 6\text{SiO}_2 + 6\text{K}_2\text{O}$$

$$\text{Ksp} + \text{Ca rich plag.} + \text{H}_2\text{O} \rightarrow \text{Na rich plag.} + \text{SiO}_2 + \text{muscovite} + \text{K}_2\text{O}$$

(2) the Schwantke model proposes that Ca in a high temperature alkali feldspar occurs in a hypothetical high silica Schwantke molecule $\text{CaAl}_2\text{SiO}_{16}$,
entailing the presence of a vacant cation site for each \( \text{Ca}^{2+} \) ion. On cooling the Scwantke molecule is converted to An and the vacant cation sites are eliminated freeing Si to combine with \( \text{O}_2 \) to form quartz. Plagioclase exsolution occurs to form perthite and myrmekite grows at suitable feldspar/feldspar boundaries.

\[
x \text{KAlSi}_3\text{O}_8 \\
x \text{NaAlSi}_3\text{O}_8 \\
\rightarrow x \text{KAlSi}_3\text{O}_8 + z \text{NaAlSi}_3\text{O}_8 + z \text{CaAl}_2\text{Si}_2\text{O}_8 + 4z \text{SiO}_2 \\
z \text{Ca}^{2+} (\text{AlSi}_3\text{O}_8)_2
\]

where \( \text{Ca}^{2+} \) = vacant cation sites, \( x > y >> z \)

Simpson (1985) produced a model which combined the above hypotheses for a non-hydrostatically stressed orthoclase grain. This involves the preferential movement of cation vacancies towards the instantaneous shortening direction which sets up a charge imbalance unless cations move at similar rates. The \( \text{Na}^+ \) radius is much less than \( \text{K}^+ \) and diffuses more rapidly and myrmekite production occurs adjacent to the shortening direction. This model was dismissed by Simpson and Wintsch (1989) because this lattice diffusion model does not adequately account for the Ca content in the newly formed plagioclase. Instead Simpson and Wintsch (1989) proposed a model in which the alkali feldspar megacrysts act as stress risers in a non-hydrostatic stress field. Under low amphibolite conditions the concentrations of elastic strain energy and strain hardening associated with immobile dislocation tangles is likely to occur at alkali feldspar grain boundaries. The composition of the ambient aqueous fluid present would be buffered in unstrained grain boundaries. Along the strained boundaries the equilibrium is destroyed and a replacement reaction occurs (Fig 6.4). This replacement of alkali feldspar by plagioclase is accompanied by a 10% volume reduction which relieves the high normal stress by a process resembling diffusive mass transfer or ‘incongruent’ pressure solution (Beach 1982). This model favours the Becke (1905) replacement model
muscovite formation in fine grained shear plane

Kspar replacement by plag.

Ca\(^{++}\) Na\(^{+}\)

Ca\(^{+}\) Na\(^{-}\)

possible plag. replacement by Kspar

modified from Simpson and Wintsch (1989)
for the formation of strain related myrmekites and can be viewed as a cation exchange reaction.

In the OMG (thin sections) possible sources for the cations (Na\(^+\) & Ca\(^{2+}\)) necessary for the replacement reaction and sinks into which the K\(^+\) cations produced are deposited have been identified. Simpson and Wintsch (1989) suggest that there are several possible sources and sinks:-

(1) Alkali feldspar replacement of plagioclase may occur in low strain ‘tail’ regions of the augen because of the rapid migration of K\(^+\) ions. This increases the K\(^+\)/Na\(^+\) activity ratio and the fluid precipates alkali feldspar, and thus these regions act as possible sources for Ca\(^{2+}\) and Na\(^+\) and a sink for K\(^+\) and can result in features such as myrmekite enclosed by alkali feldspar (Plate 6.24).

(2) Plagioclase is dynamically recrystallized to a lower temperature more sodic composition thus releasing Ca\(^{2+}\) cations.

(3) K\(^+\) cations may take part in other reactions such as the formation of muscovite (equation 6.3).

Textural evidence for alkali feldspar megacrysts overgrowing and including their matrix minerals at grain margins provides evidence that process (1) may have operated (Plate 6.25). Microprobe data presented in the next section indicates that process (2) may also have occurred. Simpson and Wintsch (1989) found little evidence in their data which suggested that process (3) operated, however several thin sections taken from the OMG display evidence for the production of muscovite relating to the deformation of specific alkali feldspar megacrysts (Plate 6.25).

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Plate 6.24.

Plagioclase and quartz inclusions in an orthoclase crystal. This has formed in the part of the crystal which is likely to be orientated parallel to the $1 + e_1$ direction in this sample. Field of view 4mm.

Plate 6.25.

An alkali feldspar crystal which displays myrmekite pools in the part of the crystal which at right angles to the maximum principle stress direction. Muscovite is formed in the shear band immediately adjacent to the large myrmekite pool. Field of view 4mm
6.6 Geothermometry using the two feldspar geothermometer

6.6.1 Geothermometry of the main foliation

Polished thin sections taken from seven samples of the OMG were analysed to establish the albite contents of both alkali feldspar and plagioclase in an attempt to apply the two feldspar geothermometer of Whitney and Stormer (1977). The analyses were carried out on orientated thin sections cut perpendicular to the foliation and parallel to the lineation. Electron microprobe analysis were performed at the University of Durham, Geological Sciences on a Cambridge Instruments Geoscan Microanalyser MkII fitted with a Link energy dispersive x-ray analyser. Chemical analysis were calculated using natural mineral standards and included ZAF correction.

Whitney and Stormer (1977) modified the two feldspar geothermometer of Stormer (1975) based on the partitioning of the albite component between plagioclase and alkali feldspar as originally suggested by Barth (1934). The geothermometer suggested by Stormer (1975) is based on thermodynamic parameters for the highly disordered (high temperature) sanidine present in volcanic suites. Whitney and Stormer (1977) produced determinative curves for the low temperature form of alkali feldspar (microcline ) and stated that variations in the structural state of alkali feldspar could shift the calculated temperature by as much as 100° C. Granitic plutonic rocks however may display a certain degree of ordering in the alkali feldspar and orthoclase will be produced for which the shift will be less than 100° C. The Stormer (1977) sanidine model is likely to give temperatures which may be up to 50° C too low.

The average albite contents from analysis of alkali feldspar and plagioclase in the seven thin sections which were analysed is presented in Table 6.1. These analyses were carried out towards the centres of grains in an effort to provide
<table>
<thead>
<tr>
<th>Sample</th>
<th>mineral</th>
<th>Albit</th>
<th>σ</th>
<th>no.</th>
<th>Temp.</th>
</tr>
</thead>
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<td>6</td>
<td>545</td>
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<td>550</td>
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<tr>
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<td>540</td>
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<td>4</td>
<td>540</td>
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<td>Plag.</td>
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<td>0.8</td>
<td>8</td>
<td>500</td>
</tr>
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<td>0.6</td>
<td>4</td>
<td>500</td>
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<td>1.2</td>
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<td>500</td>
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<tr>
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<td>Ksp.</td>
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<td>1.8</td>
<td>4</td>
<td>545</td>
</tr>
</tbody>
</table>

no. = number of analyses, Temperatures in ° C
Plag. = plagioclase, Ksp. = alkali feldspar

Table 6.1
an estimated temperature on the recrystallization of plagioclase and alkali feldspar. The central regions are likely to preserve older microstructure and textures compared with grain boundaries, along which later deformation is localized. The results are plotted on three determinative curves from Stormer (1977) and Whitney and Stormer (1977) (Fig 6.5). If it is assumed that the Stormer plots underestimates the temperature by up to 50° C, and the Whitney and Stormer curves overestimates the temperature by up to 50° C then the final result is the same for both plots at 10Kbars pressure. The emplacement depth of the OMG has been assumed to be approximately 6kbars so the data on this plot provide a minimum temperature for the feldspar recrystallization.

Brown and Parsons (1981) present what they consider to be a more satisfactory form of the geothermometer. The OMG data, when plotted on these determinative curves produce temperatures which are 40° C lower than those from Whitney and Stormer (above). The relative temperature differences are maintained however, particularly that between the main foliation development and the shear band cleavages.

These estimates assume that the both alkali feldspar and plagioclase were in equilibrium with the aqueous fluid at the time of recrystallization, which may not have been the case in a granitoid undergoing deformation. The results indicate a temperature of 540-550° C for the granodiorites and 500° C for the granites. These temperatures are approximately 100° C below the likely solidus temperature of these rocks.

6.6.2 Geothermometry of the mylonitic foliation

An attempt was made to estimate the temperature at which the main shear bands were formed. As shown in the above sections this may have taken place by a process involving myrmekite formation followed by recrystallization of plagioclase and quartz in fine grained shear bands which are further deformed
in localized zones. The myrmekite formation at high stress sites, which marked
the onset of this localization, was analysed to see if there was a detectable
temperature drop which could explain the change in deformation mechanisms
which is thought to occur at the onset of the shear band formation. To check
that this was a real effect and not an artefact caused by more albitic rich zoning
towards the margins of plagioclase; crystals were analysed at the centre and
grain margin and the results compared with that from the myrmekite zones.

The results are presented in Table 6.2 and the spots analysed are indicated
on Plate 6.26. Due to limited time for analysis the data is restricted to that
collected from a single thin section, however the results appear to be consistent
with what would be expected. The plagioclase compositions can be divided
into two categories. The plagioclase grain centres (Pl1) have a composition
which averages An30. This is zoned at the margins to An24 (Pl2). In the
myrmekite and recrystallized plagioclase the average composition is An14.
This gives a temperature estimate of 480° C for the shear bands, provided
feldspar equilibrium was attained. This result is consistent with the idea that
the shear band localization formed slightly down temperature from the main
foliation development and continued to localize as temperature dropped to
form the mylonite zones.

The data obtained by using the two feldspar geothermometer produces
results which are consistent with that predicted from the microstructure. The
absolute temperatures may not be accurate due to:-

(1) The problems of estimating the pressure at which the microstructure
was forming.

(2) Uncertainties in the structural state of alkali feldspar.

(3) The inability of the feldspars to maintain equilibrium during evolution
of the microstructure.

The relative values of the temperatures attributed to the different mi-
Plate 6.26.

This is part of the polished thin section of granodiorite which was analysed using the microprobe to provide geothermometry data. The approximate position of the spots which were probed is marked. Field of view 4mm.
Table 6.2

<table>
<thead>
<tr>
<th>Sample</th>
<th>An σ</th>
<th>Alb σ</th>
<th>Orth σ</th>
<th>no.</th>
<th>Temp.</th>
</tr>
</thead>
<tbody>
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<td>68.0</td>
<td>2.1</td>
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<tr>
<td>Pl2</td>
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<td>75.1</td>
<td>2.8</td>
<td>1.3</td>
</tr>
<tr>
<td>Pl3</td>
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<td>81.2</td>
<td>2.1</td>
<td>1.3</td>
</tr>
<tr>
<td>Kspar</td>
<td>0.7</td>
<td>0.1</td>
<td>9.1</td>
<td>1.1</td>
<td>1.1</td>
</tr>
</tbody>
</table>

no. = number of analyses, Temperatures in °C.

Pl1 = centre of plagioclase
Pl2 = zoned margin of plagioclase
Pl3 = recrystallized plagioclase in mylonitic foliation
crostructures may however be of some use in deciphering the microstructural evolution of a highly strained granite foliation as demonstrated above.

6.7 Summary and discussion

6.7.1 Environmental conditions during the textural evolution

The textural information from the OMG samples provide evidence for the operation of crystal plastic deformation mechanisms. In particular the deformation of feldspar and quartz during the main foliation development produced characteristic microstructures which indicate that they have deformed by dislocation creep processes. Simpson (1985) described the microstructural features in the different granitic components which characterized deformation at different metamorphic grades. Comparison of these features with those of the OMG show that the main foliation developed within the lower amphibolite facies.

Temperatures calculated from the two feldspar geothermometer indicate temperatures of 540-550°C for the formation of the main foliation which is approximately 100°C below the estimated solidus temperatures for granodiorite and tonalite. The formation of structures which predate the main foliation and display extreme ductility in feldspar indicate that deformation may have continued throughout the sub-solidus history. In the Ox Mountains it can be demonstrated (Chapter 2) that the transpressional sinistral shear zone was active prior to intrusion of the OMG and may well have been active during intrusion (Chapter 7). It is considered highly probable that this shear zone continued to deform the granitoids as they cooled through solidus temperatures to produce the preserved microstructures and textures at amphibolite grade temperatures and pressures.

The formation of myrmekite at sites of high normal stress on alkali feldspar
megacrysts occurs widely throughout the OMG samples and also indicates lower amphibolite facies deformation (Simpson 1985, Simpson and Wintsch 1989) and this is supported by temperatures calculated from the geothermometer. Hibbard (1979, 1987) used the presence of myrmekite to indicate the presence of aqueous-rich fluids during deformation prior to complete crystallization of the magma. The present author agrees with Simpson and Wintsch (1989) that myrmekite formation at points of high normal stress in a foliation cannot be attributed to the crystallization of late fluids. In a non-hydrostatic stress field residual magma should crystallize in areas of low normal stress or 'strain shadow regions' adjacent to phenocrysts. Simpson and Wintsch (1989) proceed to suggest that the development of myrmekite may take place after a considerable time interval between crystallization and deformation and should not be used solely to infer late stage magmatic deformation. In the OMG there is additional evidence which supports the operation of the shear zone throughout the crystallization and sub-solidus histories and the presence of strain induced myrmekite is used to constrain the environmental conditions during deformation at a particular stage in the sub-solidus textural evolution.

6.7.2 Granitoid rheology at lower amphibolite facies conditions

The most important textural feature in many of the thin sections taken from the OMG is the formation of the S-C fabric, or shear band cleavage, which deforms the main foliation (Chapter 4). In the OMG this shear band cleavage development appears to be controlled by the formation of myrmekite at the margins of alkali feldspar megacrysts.

This reaction is essentially a cation exchange reaction which is enhanced by the build up of dislocations in the parts of the crystals which face the principal normal stress direction. At higher temperatures the dislocation build up is relieved by recovery accommodated dislocation creep, which produces sub-grain formation and rotation (Tullis & Yund 1985). As temperature drops
during cooling, the recovery processes become less dominant and the diffusion of cations is sufficiently rapid to permit alkali feldspar replacement by plagioclase and quartz to form myrmekite. These plagioclase and quartz crystals are further deformed and may undergo recrystallization accommodated dislocation creep, which forms fine grain size zones in which large strains may be accumulated by grain boundary sliding.

If grain boundary sliding becomes dominant the S-C fabric is characterized by relict megacrysts aligned along S planes and the shear planes (or C planes) display large displacements. If recrystallization accommodated dislocation creep continues to operate the overall grain size is reduced to a fine/medium grained mylonite with relict plagioclase grains which may rotate in the mylonitic groundmass, or fracture if stress build up becomes too great.

This change in deformation mechanisms for feldspar has been predicted by Tullis and Yund (1985, 1987) who suggested that it may be a mechanism for shear zone formation. This model can be combined with the Simpson and Wintsch (1989) model for deformation enhanced myrmekite production. The textures present in the OMG samples support the operation of an integrated model in which a change in deformation mechanisms in response to lowering temperature or pressure promotes a deformation enhanced replacement reaction which localized the non-coaxial strain, causing shear band formation and ultimately the production of mylonites.

The rheological behaviour of crystalline granitic rocks, at amphibolite facies, is controlled by the deformation mechanisms which operate in both alkali feldspar and plagioclase. In particular the ability of feldspar to undergo recovery processes dictates whether deformation localizes or remains steady state. In comparison, at lower metamorphic grades feldspar behaves passively or fractures under high stress conditions. In this case, steady state deformation is controlled by quartz and biotite microstructural characteristics.
CHAPTER 7

THE EMLACEMENT AND DEFORMATION OF THE OMIC

The purpose of this chapter is to develop a model for the emplacement of the Ox Mountains Igneous Complex. The emplacement siting is discussed followed by the development of a model for the emplacement mechanism and subsequent deformation of the OMG and the adamellite plutons. The emplacement mechanism is discussed and compared with that of granites in other shear zones. Finally, the nature and origin of the granitic magmas are discussed and their relationship to regional tectonic models and broader scale continental lithospheric processes.

7.1 Emplacement of the Ox Mountains Granodiorite in a pull-apart or releasing bend structure?

A structural, kinematic and metamorphic model for the evolution of the Ox Mountains Inlier has been outlined in chapter 2. The OMG occupies a position, within the braided high shear strain system, along the trace of the major F₃ antiformal structure, the granite emplacement being syn-kinematic with respect to D₃ (see section 2.6.1). Hence the pluton has been emplaced during major contractional (compressional) deformation in the country rocks. In Chapter 4 a forceful emplacement which could explain this relationship has been discussed and ruled out by the strain data. This implies that a type of passive emplacement mechanism was in operation which permitted the granite intrusion in its present position.

Evidence from the contact zones on the NW and SE suggests that intrusion took place by the emplacement of sheets, on various scales, sub-parallel to
the regional S\textsubscript{3} fabric strike but oblique to its dip (section 2.6.2). The general sharp sheeted contacts between the various components of the pluton (seen most clearly between granodiorite and tonalite) suggests that an essentially brittle dyking and sheeting mechanism occurred during pluton construction. The early muscovite granite intrusions along the NW contact may also have been intruded in a similar manner, but these have been subsequently modified by amphibolite facies ductile shearing and in some cases by the lower temperature deformation associated with movements on the Knockaskibbole Fault system (see section 4.10) leading to difficulty in the determination of their intrusive mechanism. With the later granodiorites there is also internal sheeting involving slightly differing compositions and textures. This is best displayed by the biotite granite and type 1 granodiorite sheeting, e.g. Pontoon (see 3.4.3) and also on a large scale, along the SE contact where a 500m wide, uniform sheet of type 2 granodiorite was intruded.

The tonalite contact relationships, described in detail in Chapter 3, are the most clearly displayed in the OMG. Their emplacement model is considered in detail below as a guide to the overall emplacement of the sheeted pluton. In general the tonalites form elongate bodies, sub-parallel to the main pluton contacts and the main shear zone fabric. There are three emplacement models which can be suggested for these rocks.

(1) The tonalites are syn-magmatic with the granodiorites and their elongate shape is a result of subsequent deformation in the shear zone.

(2) The tonalite was intruded into relatively solid granodiorite, it crystallized and was subsequently sheared and flattened to produce elongate, boudin shaped bodies.

(3) The tonalites were intruded as a series of dykes or sheets into relatively solid, high temperature granodiorite sub-parallel to, and syn-kinematically with the main shear zone fabric.
Option (1) is ruled out because the tonalite contact relations do not indicate the former presence of liquid/liquid contacts such as sheared lobate margins and zoned contacts where magma mingling has taken place.

Option (2) would suggest that the tonalites were intruded prior to deformation and have been subsequently flattened and aligned by either pure shear or simple shear or both. In general we can rule out this option because the shape ratios of internal dykes within each tonalite body is far greater than the shape ratios of the bodies themselves. For example at Pontoon Dancehall (G212 034) dyke shape ratios are in excess of 15:1 whereas xenolith shape ratios are approximately 9:1.

(Option (3) seems to explain the tonalite contacts more successfully and the tonalite zones can be viewed essentially as a series of dykes or sheets concentrated in linear belts sub-parallel to the pluton contacts. A possible problem remains in that the overall shape ratio of the tonalite zones is about 8-10 which after removal of the ductile strain component leaves the tonalite zones as almost circular or ovoid bodies, however individual dyke shape ratios are much higher.

This leaves option (3) as the most likely model.

In summary the evidence above suggests that a brittle dyking mechanism was responsible for the intrusion of the tonalites. This is likely to have been applicable to the other units of the OMG: the early muscovite granites were intruded as a series of sheets; the granodiorite intruded as sheets along the country rock contacts and was itself sheeted internally with batches of slightly different composition. The diorites may have been intruded as sheets but differ in that they are likely to have been emplaced in a magmatic host.

The intrusion of dykes or veins is largely thought to occur at a high angle to shear-zone boundaries. Durney & Ramsay (1973) predict intrusion at 45° to the shear zone walls (90° to the $1 + e_1$ direction of the finite strain ellipsoid)
Fig. 7.1

Initial orientation 45°

shear zone cleavage at a high angle to the dyke contacts

Predicted dyke geometry if intruded at 45° to a simple shear zone
Fig. 7.2

(a) The orientation of dykes in a sinistral releasing bend structure from Sanderson & Marchini (1984)

(b) Orientation of dykes in a sinistral pull-apart structure
for a simple shear type shear zone. The dykes become rotated by further ductile strain and develop a sigmoidal shape. If dyke intrusion is accepted as a valid intrusion mechanism for the OMG then there should be evidence for widespread rotation of those dykes. Figure 7.1 shows the amount of rotation predicted by assuming that the average xenolith shape ratio is a product of simple shear. It is apparent that very high shear strain values are required to rotate the dykes close to parallelism with the shear zone fabric. It would be expected that since there are strain variations across the pluton, rotation by simple shear of the dyke contacts should form large scale sigmoidal strike variations in which dyke contacts in hinge zones are at a high angle to the main foliation. Large scale examples of this phenomena have not been recognised in the OMG during this study and indeed the zone 1 tonalite orientation would have had to rotate through the shear plane orientation to reach its present orientation which is theoretically impossible.

The present orientation of these intrusive structures can not be attributed to their formation in a sinistral shear zone releasing bend segment or a pull-apart structure followed by rotation and flattening. Intrusion in either structure is likely to produce sheeting at a higher angle to the shear zone fabric (Sanderson & Marchini 1984) than would be predicted for simple shear (Fig 7.2). The sheet geometry in the contact zones and the internal contacts suggest that the emplacement mechanism was not one which formed as a result of deformation in a transcurrent shear zone. It is necessary to consider the shear zone as a transpressional structure to acquire a better understanding of how the OMG was emplaced.
7.2 The emplacement mechanism

7.2.1 The emplacement related structures

The emplacement related structures have been described in Chapter 2 and consist of a series of brittle, gently inclined overthrust structures which deform the main $S_3$ fabric and are in turn overprinted by a continuation of the $D_3$ shearing. The displacement on these structures in the southern contact zone is consistently up towards the NW and up towards the SE in the northern contact zone. Some of these brittle thrust structures also contain sheets of the main granodiorite. At Burren hill (see section 2.6.2) it can be demonstrated that the granodiorite is intruded along thrust planes while they were actively moving and indeed, the magma is likely to have been responsible for the propagation of the thrust structures. The strain rate associated with the accession of the magma may have exceeded the ductile strain rate and, hence caused a transition to brittle failure.

The introduction of the magma along the thrust planes implies that the footwalls and hangingwalls are moved apart in a plane at right angles to the thrust azimuth. The magma has therefore produced a vertical extension which may also be a vertical dilation. A vertical extension would not be expected in a transcurrent shear zone but may occur in a transpressional shear zone system (Fig. 7.3). Given the vertical extension associated with the intrusion events, it seems likely that the intersection between the sheets in the contact zones and the $S_3$ cleavage is sub-parallel to the $1 + e_2$ direction. This plunges gently to the NE or SW in all areas and $1 + e_1$ is likely to be sub-vertical. (Fig. 7.4). The main shear zone deformation which extensively deformed both the igneous and metasediments lithologies after the intrusion event has a well developed, sub-horizontal stretching lineation formed in an intense, planar fabric. This indicates that the during shear zone deformation,
Vertical extension in a zone of transpression

from Sanderson and Marchini (1984)
Fig. 7.4

Orientation of the finite strain ellipsoid before and after intrusion of the OMG

Orientation of the finite strain ellipsoid during intrusion of the OMG
both before and after intrusion, that $1 + e_1$ was sub-horizontal. Predominantly transcurrent deformation was interrupted and modified by a component of vertical extension at the time of intrusion.

### 7.2.2 The Tonalite dyke zones

The original intrusive geometry of the dykes, after removal of the ductile strain is considered to form two sets of moderately inclined sheets (see section 4.5). The original dip is not known, but likely to be 40-50° NW in the northern part of the pluton and 40-50° SE in the southern part (see Figure 4.8). If these tonalite sheets are formed in a conjugate set then it may be possible to determine the orientation of the stress system during their intrusion. The angle between conjugate shear zones and the maximum principle stress ($\sigma_1$) is known to vary with depth (Ramsay 1983, Kligfield et al. 1984). At depths greater than 10Km and amphibolite or granulite facies metamorphism $\sigma_2$ is thought to vary between 120° and 90° whereas in the upper parts of the crust it is thought to 60°. The uncertainty in the variation with depth mean that it is not possible to accurately determine the stress field orientation for the tonalite dykes. It is probable however, that the intersection of the conjugate dykes corresponds with the intermediate principle stress ($\sigma_2$) (Fig. 7.5).

A transcurrent shear zone system is likely to have a stress system in which $\sigma_2$ is vertical. The stress systems associated with transpressional systems are difficult to predict in detail, however the Ox Mountains Shear Zone which has a sub-horizontal stretching lineation ($1 + e_1$ of the finite strain ellipsoid for the ductile shearing) is likely to have a similar stress system to a transcurrent shear zone. The evidence from the tonalites suggests that the stress system changed from one in which $\sigma_2$ was vertical to one in which $\sigma_2$ was horizontal during the intrusion process (Fig. 7.5).

The evidence discussed in the above sections suggest that the intrusion of the OMG occurred in an area of vertical extension and dilation. This is
Possible orientation of the stress system during tonalite intrusion
formed in the centre of the shear zone system along the trace of the major antiformal axis. This region is likely to undergo maximum contraction in the transpressional shear zone. This may be accommodated by upward movement of the central block of the shear zone causing vertical extension. During this process differential movement may occur on the high strain zones on either side of the central block which contains the antiformal structure. Relative upward movement on the outside parts of the shear zone and relative downward movement of the inner part creates a zone of vertical dilation in the antiformal hinge region (Fig. 7.6). The overall emplacement model resembles a large scale flexural slip model in which the formation of 'a saddle reef' has occurred.

The detailed emplacement mechanism however, is that the pluton is constructed by episodic sheeting events with the intrusion of a small volume of muscovite granite followed by the more major units; biotite granite, granodiorite and tonalite. The final event was the intrusion of gently inclined sheets of microgranite These were intruded in a similar orientation to the main units but contain a weak fabric and their contacts have not been steepened by the ductile strain. Their orientation indicates that at the late stages of the intrusive phase the intrusive stress system was similar to that during the earlier intrusive events. This episodic sheeting may have occurred in response to pulses in the relative rates of transpression and strike-slip movements causing episodic tapping of a magma chamber at depth which produced batches of magma with different compositions.

The introduction of the granitic magma into the shear zone system creates an interruption in the overall ductile strain history. The accession of the magma may occur rapidly, producing a brittle, high strain rate environment. This may reduce the differential stress which could lead to a situation in which large volumes of magma are emplaced probably accompanied by rapid uplift of the central block of the shear zone. Following emplacement the granitoid
The siting of the OMG in the central part of the transpressional shear zone.
cools rapidly and crystallizes changing its rheological response rapidly as it passes through the critical melt percentage (see section 3.10.3). During the last stages of crystallization the melt may become trapped in pools at grain boundaries. Pressure build up in these pools may cause a rupturing of the crystal framework which may account for the presence of the early discrete shears which are overprinted by the main fabrics. Upon full crystallization the granite deforms with a plastic rheology similar to the surrounding country rocks and the shear zone reverts to its normal ductile pervasive shearing. This indicates that the stress system and principle strain axes have returned to their normal transpressive shear zone position. The post-intrusive structural evolution in the OMG indicates that sinistral shearing evolved via a series of discrete shears, through a pervasive ductile main fabric and S-C cleavage development into localized zones of shear with antithetic dextral shear zones. This is consistent with a model of continued ductile sinistral transpressive shearing on a cooling crystalline pluton (see chapter 4). This produced the final steepening of the main S3 fabrics and both country rock/granodiorite contacts and internal igneous contacts.

7.3 The emplacement of the adamellites

Details are given in Chapter 5 concerning the adamellite field relations and in particular, the timing of intrusion relative to the deformation chronology. The regional D4 structures are cross-cut by both adamellites which correlates with their c400ma age date.

Significantly the site of emplacement of the LTA and ELA is adjacent to a releasing bend in the splay of the Glennawoo shear zone (see Chapter 2). There is little direct evidence in the metasediments adjacent to the adamellites for a post-D4 reactivation of this structure, however both adamellites display
their most intense foliation and sinistral S-C fabric development on the margin adjacent to the Glennawoo shear zone. There is evidence for a contact parallel magmatic state fabric (modified by sub-solidus sinistral shear in the LTA) which may represent an original emplacement related foliation.

There is no evidence of sheeting in these plutons, either internally or at their margins, and although the exposure is not good it seems clear that these plutons have been each emplaced in a single magmatic batch. It is not possible to closely constrain an emplacement model for the adamellites, but they may have been emplaced in a type of extensional cavity created by late discrete movements on the Glennawoo shear zone. The emplacement depth of 10Km may explain why they have fundamentally different emplacement style to the OMG as extensional cavities, e.g. pull aparts, are more likely to exist at higher levels in the crust. Figure 7.7 illustrates the possible shape and an emplacement model for the LTA. The floor contact preserved under this pluton (see 5.2.1) indicates that this adamellite was emplaced as a thin sheet. The footwall to this structure is likely to have remained attached to the sidewall of the Glennawoo shear zone whereas the hanging wall became detached. Renewed sinistral movement at c400Ma resulted in the footwall sliding away from the hanging wall creating an extensional cavity into which the granitic magma ballooned or was drawn. The geometry of the structural information for the ELA is displayed in Figure 7.8, with a possible three dimensional shape. The granitic magma in this case is likely to have ballooned into the low pressure region in the releasing bend adjacent to the Glennawoo shear zone.

Both plutons are considered to have been emplaced in response to reactivated sinistral shear on the Glennawoo shear zone which may have created some type of extensional cavity or low pressure area into which granitic magma was forced or drawn into. This was subsequently deformed by sinistral shear
c400Ma rectivation of the shear zone produces an extensional cavity in which granite balloons or is drawn. This cavity is formed beside the solid OMG.

The LTA displays a strong solid state foliation at the margins with sinistral shear bands and a late dextral semi-brittle deformation.
c.400 Ma reactivation of the shear zone creates an extensional cavity into which the granite is emplaced.

ELA is emplaced in a releasing bend structure.

Easkey Lough shear zone

ELA displays a solid state foliation with sinistral shear bands and late semi brittle dextral deformation.
and then overprinted by late, dextral, semi-brittle shearing in both plutons. The reactivated sinistral event may also have had a slight transpressional component which would partly explain the foliation patterns which have been preserved.

7.4 Discussion of emplacement mechanism

There have been very few examples in the literature in which granite emplacement by sheeting or dyking has been described in detail. In the British Isles, the only other example which begs comparison is the Main Dongal Granite. This pluton has been mapped in detail by Pitcher and Read (1958) and co-workers, who produced a model in which lateral magmatic wedging emplaced large sheets of granite horizontally from the NE-SW and this was accommodated by dilation and lateral distension of the country rocks. Hutton (1982b) proposed that the Main Donegal Granite is emplaced entirely within a sinistral shear zone which dilated by bending in response to lock up at the shear zone tip, thus drawing granitic magma into the shear zone. An important part of this model is the recognition that the pluton contacts are highly strained shear zones, relative to the central parts of the pluton.

Hutton (1982b) did not attempt to relate the geometry of the internal contacts to the stress system likely to have been operating during the syntectonic intrusion. Examination of Pitcher and Read's (1958) map of the Main Donegal Granite shows that granite sheets are intruded up to 1.4km outside the contact but are most common up to 500m outside both the N and S contacts. Most internal contacts appear aligned sub-parallel to the pluton long axis (see Plate 4.6). The internal strain magnitudes average approximately 5.0, as measured from xenolith X/Z measurements, which is unlikely to have rotated contacts into parallelism with the shear plane. It is suggested
that a similar mechanism to that suggested for the OMG may have operated in the Main Donegal Granite and the main shear zone deformation localized along the outer margins because it took place at higher levels in the crust where the pluton was able to cool much more rapidly.

The work of Shaw (1980) and Spera (1980) established that fracture mechanics are a valid mechanism for magma transport, however there have been very few examples where actual emplacement of granitic magma has been attributed to the operation of brittle processes.

Castro (1986) proposed that the Central Extremadura Batholith was emplaced by extensional fracturing. These fractures developed parallel to a regional horizontal shortening associated with high grade metamorphism at depth. In his model these fractures propagated to the mantle causing the ascent of basic magmas which extensively melted the base of the crust, creating granodiorite melts which were intruded via extensional fractures. The fractures form at 45° to an E-W shear couple in the basement and rotated from the extensional field to the compressional field of the strain ellipsoid where they underwent transverse shortening. Castro (1986) did not have other evidence to support an E-W shear couple other than the necessity to have extensional fractures at 45° to it. This author's work in the OMG has suggested that under certain conditions extensional shears do not necessarily form at the classic 45°, but may be formed almost parallel to shear zones.

7.5 Granitoid rocks and shear zones

A review of recent literature shows a number of examples of granitoids which have been emplaced in transcurrent shear zones. It provides an interesting insight to separate those formed in dilatational sites from those which have been emplaced in a transpressional environment. Four plutons emplaced
in shear zones with an dilational component are:- Mortagne (Guineberteau et al. 1987), Main Donegal Granite (Hutton 1982b), Strontian (Hutton 1988a) and perhaps the Leinster Batholith (Cooper & Bruck 1977 and Murphy 1987).

There are many more examples of granites emplaced in shear zones which are thought to contain a transpressional component. The emplacement invoked is usually diapirism or oblique diapirism. Examples in the Hercynian including:- the Lacronian and Pontivy leucogranite massifs (central Brittany) (Hanmer & Vigneresse 1980 and Strong & Hanmer 1981) and other plutons intruded along the South Armorican Shear Zone (Gapais & LeCorre 1980 Jegouzo 1980 and Berthe’ et al. 1979). Other Hercynian plutons emplaced in probable transpressional shear zones are; the Serra de Frieta (Reavy 1988, 1989) and the syn-kinematic leucogranite diapirs in Galicia which are intruded in regions of maximum shear zone intensity (Ponce de Leon & Choukroune 1980). The Najd fault zone in Saudi Arabia contains zones of volume loss and gain which are sub-parallel to the shear zone boundaries (Davies 1982). Granite intrusion occurs by diapirism mainly in the zones of volume gain. In the Gander Zone of Newfoundland an Acadian ductile transpressional shear zone contains several granitic diapirs (Hanmer 1981). The Caledonian oblique diapir, Criffel, may be related to sinistral oblique movements (Soper & Hutton 1984).

It is apparent from the above that granites appear to be more common in transpressional shear zones rather than in pure strike slip systems. Syn-kinematic plutons are also commonly associated with oblique thrust type shear zones. eg, plutons emplaced in the late Mesozoic to early Cretaceous events in the Eastern Cordilleran metamorphic belt in the western USA (De Witt 1980, Anderson & Rowley 1981 and Haxel et al. 1984). Granites are also common in more orthogonal collision zones such as the Himalayan leucogranites, (LeFort 1988) and above major subduction zones such as the Peruvian
Andes. (Pitcher & Bussell 1977). Pitcher (1978) emphasized the structural control of deep seated lineaments in the crystalline basement on the siting of the Coastal Batholith of Peru. Large differential vertical uplift occurred by movement on these lineaments probably in response to the intrusion of granitic magmas. Pitcher (1978) suggested that the main phase of granitoid emplacement occurred during periods in which there was no compression or possible extension. Pitcher and Bussell (1977) proposed that the fault patterns, which controlled pluton emplacement, were a result of the interplay between a regional compressive regime and one of vertical uplift. They also noted the presence, at a deeper level, of forcefully emplaced diapiric plutons which have formed in more ductile environments. They acknowledged that a hydraulic fracturing mechanism could possibly explain the formation of plutonic contacts and dyke fissures but preferred the pre-existing structural lineament model for the overall batholith construction. A hydraulic fracturing model is similar to that developed during this study and on a much larger scale the creation and ascent of large volumes of granitic magmas may locally reverse the regional compressive stress field. This may occur at lower and middle crustal levels to permit the ascension of granitic magmas to higher levels where they are emplaced as large cauldrons.

Granite plutons associated with true extensional regimes are rare, although one such example is the late Proterozoic Rapakivi Granites of S. Greenland Becker & Brown (1985) and Hutton et al. (in press).

7.5.1 Source regions for granitic magmas

The varied nature of granitic magmas reflects the fact that they can originate by a number of different processes and in different parts of the continental lithosphere. Fyfe (1987) suggests six different melting sites above a subduction zone and five in a continental collision zone. It is not an object of this thesis to develop a model for the formation and segregation of the granitic
melts which form the OMG. However, it is useful to speculate on the nature and origin of the granitoid rocks in the Ox Mountains in an effort to further constrain the regional significance of the Ox Mountains Inlier. Without geochemical and isotopic data it is unwise to advocate a particular source region or mechanism by which the granodiorites melts are formed and this study is confined to the use of petrological information in comparison with that known from other tectonic environments.

There are four major granite forming environments:

(1) In Oceanic Island Arcs, extreme fractionation of basic magmas in the crust will produce small volumes of plagiogranites. This is the M-type granite of Pitcher (1984).

(2) The high SiO₂ granites which are leucocratic monzo-granites and S-type granitoids of the Hercynian and the High Himalayan granites. These are typical of crustal thickening scenarios and the melts are likely to form in low pressure, high temperature metamorphic regions where anatexis may occur at 12-15km depth. England and Thompson (1986) suggested that the temperature rise due to crustal thickening alone can lead to melting and this combined with enhanced fluid flow in shear zones can allow formation of large scale granite batholiths (Reavy 1989). There is also likely to be a sub-crustal contribution to the thermal structure because the continental lithosphere must also thicken during continental collision (Houseman et al. 1981).

(3) Granitic rocks which form magmatic arcs adjacent to subduction zones, e.g. the Peruvian Andes and the Sierra Nevada, are not true granites but predominantly tonalitic in composition and are associated with gabbros. These form large, linear batholiths and are associated with large volumes of andesitic and dacitic volcanics. These granitoids may form by remelting of underplated diorite which is formed by metasomatism or partial melting of mantle overlying the subducting slab (Wyllie 1983). Fluids released by prograde metamor-
phism of the subducting oceanic crust and lithosphere can promote melting and metasomatism of both the slab and the overlying mantle, producing a variety of compositions by hybridization and mixing.

(4) Granites may form in association with post-orogenic uplift and strike-slip faulting as in the Late Caledonian granites of Britain and Ireland. These are predominantly granites and granodiorites which are associated with basic magmas. They may have formed by adiabatic decompression of the lower crust during uplift.

On purely petrographic terms the OMG rocks are more closely akin with either 3 or 4 (see chapter 3) although a precise classification would require the use of isotopic methods to quantify the mantle input. The OMG have an initial \(^{87}\text{Sr}^{86}\text{Sr}\) of 0.7056 (Pankhurst 1976) which would classify the granitoids as post-Caledonian uplift I-type granites (Pitcher 1983). However, these rocks do contain more tonalite and diorite than most of the later c400ma plutons suggesting that they may have formed in slightly different environment.

7.6 Caledonian melting processes

The late Caledonian c400Ma granitoids of the British Isles are considered to have formed following a post-orogenic uplift setting (Dewey & Pankhurst 1980). However, recent high precision U-Pb dating of zircons from the Ben Vurich granite (Scotland), which was emplaced between D2 and D3 of the Dalradian block has provided an age of 590 ± 2Ma (Rodgers et al. 1989). This new dating implies an age gap of at least 190Ma between the main crustal thickening and metamorphism and the intrusion of the post-uplift granitoids. An alternative hypothesis might be that the late Caledonian granites are a product of a c400ma thermal event unconnected with Grampian deformation or uplift but associated with the widespread transcurrent shearing known to
occur at this time (Hutton 1987). A problem remains in that the OMG dated at 478±12 is older than these Late-Caledonian granites.

The OMG melts are likely to have been formed by one of the melting processes outlined above which are:

(1) Fractionation of basic rocks.

(2) Adiabatic decompression.

(3) Crustal thickening.

(1) Fractionation is ruled out because the Ox Mountains does not contain evidence that it was part of an Oceanic Island Arc enviroment. The OMG is intruded into sediments which have strong continental affinities. There is also no evidence for the large volumes of basic rocks which would be necessary to produce a pluton of the magnitude of the OMG.

(2) Adiabatic decompression has been advocated for the late Caledonian granitoids and is supported by the recognition of a widespread retrogressive event prior to their intrusion. In the Ox Mountains the LTA and ELA are both intruded after a major uplift phase indicated by widespread retrogression (see chapter 2). Indeed the end-Silurian strike-slip faulting may well have assisted that adiabatic decompression process by producing accelerated local uplift on certain blocks. This process cannot account for the OMG melts as they were intruded just after the peak of regional of regional metamorphism, prior to the uplift stage.

(3) Crustal thickening of some sort is the likely explanation for the formation of the OMG magmas. A mantle input is suggested by the petrography, the range of rock types and the relatively low initial Sr ratio. Houseman et al. (1981) suggested that during crustal thickening the underlying lithosphere must also be thickened. The thickened boundary layer between the cool lithosphere and the convecting asthenosphere may become unstable and sink to be replaced by the hotter asthenosphere. This will transfer heat to the crust
and upper mantle contributing to the regional metamorphism to produce underplated regions by partial melting of the upper mantle and crust. The time delay is estimated to be 30-50my which may be of the order of certain orogenic events.

The presence of fluids in shear zones is likely to be another major contributory factor in the formation of granites in regions of thickened crust. Shear zones are known to be major conduits for fluids, although the depths to which they might penetrate are not known accurately. In subduction zones Fyfe (1988) estimates that volatiles could be carried to 100km depth in the subducting slab. The release of volatiles during prograde metamorphism will provide episodic bursts of fluids which may induce partial melting of the mantle lithosphere.

7.6.1 Crustal thickening in the Ox Mountains

Deformation in the Ox Mountains is thought to represent that which took place in the side-wall to, and during the early history of the Highland Boundary Fault system (see chapter 2). Subsequent movements along this fault have obscured the likely cause of the crustal thickening event. The crustal thickening associated with the transpressional shear zone such as the Ox Mountains shear zone is not likely to cause a large enough perturbation of the thermal structure of the continental lithosphere to produce a widespread melting event and it is necessary to turn to structures of more regional significance to produce the necessary thickening event. There are two possible scenarios in which crustal thickening may have occurred in the Ox Mountains.

(1) Ophiolite obduction.

In this model the numerous Caledonian ophiolite fragment, e.g. Tyrone, Bay of Islands, Shetland and Quebec, which are dated at c465-490Ma belong to a large sheet (Dewey & Shackelton 1984) which has been obducted along the eastern margin of Laurentia. There are fragments of this ophiolite smeared
along the Highland Boundary Fault (Dewey & Shackelton 1984). Serpentinite adjacent to Clew Bay (County Mayo) and along the Knockaskibbole fault in the Ox Mountains are thought to be of this age. The obduction of an 15Km thick ophiolite sheet may have led to crustal depression and lithospheric thickening in a geometrically similar manner to that suggested by Houseman et al. (1981). The subsequent development of the HBF transpressional shear zone system may have further perturbed the thermal structure of the lithosphere creating small areas of partial melt in an underplated region of dioritic composition, thus forming small areas of granitoid melt such as the OMG (Fig 7.9a). One problem with this hypotheses is that the OMG, dated at c480Ma is closely related in time to the ophiolite obduction which would mean that in theory the lithospheric delamination would not have time to develop. Any possible connection may be resolved through better age dating of the OMG.

(2) Following the work of Pitcher and Berger (1972), Alsop (1987) showed that the Lough Derg Slide is a major dextral, oblique, ductile thrust which translated a 20km wide thrust sheet SE over the Lough Derg basement. These latter rocks have been correlated with the NE Ox granulites. The age of this structure is problematical but is known to post-date a retrogression of the main MP2 event which is developed throughout the Dalradian rocks of Donegal. This structure has been correlated with a similar thrust developed in the Sperrin Mountains of County Tyrone (D. Hutton pers. comm) and is likely to have created a major crustal thickening over the basement to the South. The associated lithospheric thickening may have undergone a delamination process (Fig 7.9b). The timing and kinematic link between this thickening event and the Ox Mountains has not been resolved. If they were linked however, a steep shear zone would be likely to provide a conduit for granitic magmas and may also assist the magmatic event by triggering melting of an underplated region of the crust and upper mantle. The channelling of fluids, formed during the
The ophiolite sheet depresses crust and thickens the mantle lithosphere. Transcurrent shear on HBF triggers melting of underplated region.

Crustal thickening by a major Caledonian shear zone combined with transpressional HBF deformation which provides an additional thickening generating localized melting in Fault zone.

after Houseman et al (1981)
prograde metamorphism, by the shear zone must be emphasized in relation to the causes of the melting process.
CHAPTER 8

CONCLUSIONS

8.1 Major conclusions to chapter 2

(1) The work of Jones (1989) has shown that the SW and Central Ox Mountains Inlier is composed of Dalradian metasediments and volcanics which are in tectonic contact with Pre-Caledonian basement granulites in the NE Ox Mountains. He correlated these rocks with Dalradian rocks in Central Donegal.

(2) Jones (1989) established a single deformation chronology which could be correlated throughout the SW and Central Ox Mountains Inlier. This work produced a chronology in which the tectonothermal history of the Ox Mountains could be split into five main structure or fabric forming episodes. D1 produced no major structures in the Ox Mountains but did form a fine grained alignment fabric. The only major D2 structure is the tectonic slide between the Ardvarney Formation and the Ox Mountains Sucession. In the remainder of the Ox Mountains D2 is represented by a penetrative shape and alignment fabric and associated minor F2 folds. The major deformation event in the SW and central Ox Mountains is the development of a D3 sinistral transpressional shear zone. This produces a major, upright F3 fold. Minor F3 folds are upright, close to isoclinal with their axial planes parallel to the strongly developed S3 cleavage and axes parallel to the gently NE or SW plunging stretching lineation. The steep, NE-SW striking S3 cleavage transposes the earlier fabrics in high strain mylonite zones. These zones are deformed by sinistral extensional crenulation cleavages and form a braided system of
high strain zones throughout the Ox Mountains Inlier. D₄ reflects a change in kinematics and is a series of folds with moderately inclined axial planes which indicate a dominantly vertical shortening in the Ox Mountains. There was no fabric development associated with this deformation episode. D₅ is a reactivation of the sinistral transpressional movements which deform two c400ma plutons.

(3) The main transpressional event is considered to represent early sinistral movements along the Fair Head - Clew Bay line which has been interpreted as the continuation of the Highland Boundary Fault in Ireland.

(4) The metamorphic peak in the SW and central Ox Mountains is considered to be kyanite amphibolite facies which represents a peak metamorphic temperature of 600-620° and pressure of 6-7Kbars. The peak occurred at an MP₂ stage in the deformation chronology.

(5) This study has shown that emplacement of the Ox Mountains Granodiorite occurred synchronously with the late stages of the D₃ deformation. The main ductile D₃ deformation is preserved in rafts in the pluton and continues to deform the granodiorite after it had crystallized. A narrow thermal aureole up to 300m wide produced sillimanite in pelitic wall rock lithologies. The Ox Mountains granodiorite has been dated at 478±12 on a Rb/Sr whole rock isochron.

(6) The style of deformation during the intensive events displays a marked change from pervasive ductile shearing to the development of brittle structures. This produces a series of thrusts which are associated with the granitic rocks in the main contact zones. On both the northern and southern flanks of the pluton these structures displace the country rocks up towards the centre of the pluton.
8.2 Major conclusions to Chapter 3

(1) The OMG has been divided into four major components during this study; granodiorite, tonalite, muscovite granite and diorite.

(2) The granodiorite component has been divided on a textural basis into two groups. Group 1 granodiorites are equi-granular and can be found in the northern and central parts of the pluton. They are associated with a series of biotite granites which have similar textures. The Group 1 granodiorites are intruded into the metasediments on the northern flank of the inlier as a series of sheets in the contact zone. The Group 2 granodiorite is porphyritic with large alkali feldspar phenocrysts and is the typical facies present along the southern flank of the pluton. Group 2 granodiorite forms sheets and veins in the metasediments up to 1.5km to the south of the pluton. The contact in this area is a transitional sheeted zone from granitic veins in metasediments to small metasediments rafts in the granodiorite pluton.

(3) The Ox Mountains tonalites form elongate dyke zones up to 8km long and 800m wide and are intruded into the granodiorites and biotite granites. They are exposed predominantly in the SW part of the pluton. There are three main tonalite dyke zones which are elongate sub-parallel to the long axis of the pluton. Each zone is composed of a series of tonalite bodies which are approximately aligned along strike from each other. These bodies are made up of numerous tonalite dykes which are intruded alongside each other. The boundaries to the tonalite bodies are regions of transitional sheeting between 100% tonalite dykes and the host rocks. The tonalite is generally equigranular, medium grained and composed of plagioclase, biotite and quartz.

(4) A series of isolated muscovite granites are present in the northern contact region of the pluton. These are not in contact with the main granodiorite or tonalite components and are everywhere in contact with the metasedimen-
tary wall rocks. They may have been intruded prior to the main body of the pluton because small muscovite sheets have been mapped in metasedimentary rafts in the granodiorite. These rocks are mainly composed of alkali feldspar, plagioclase, quartz and muscovite.

(5) There is a series of diorites which are present in the southern part of the pluton. These may be related to appinitic rocks which are found to the south of pluton. The contact relationships between the Group 2 granodiorites and the diorites suggests that both were liquid at the time of intrusion. The diorites are principally composed of hornblende, plagioclase and minor amounts of biotite.

(6) Minor components include pegmatites, aplites, microdiorites microgranites and quartz veins.

(7) The crystallization sequences which have been estimated for the OMG components indicate that epidote was the first mineral to crystallize followed by plagioclase, biotite, orthoclase and quartz. This does not entirely agree with the experimentally derived crystallization sequences. The presence of magmatic epidote is thought to indicate pressures during crystallization of at least 8kbars and therefore it must have formed before the final emplacement of the magma.

(8) The results of this study suggest that the OMG is constructed by the emplacement of sheets of granitic magma of differing composition. The muscovite granites form small sheets along the contact zone. The main granodiorites are intruded as sheets on both the northern and southern pluton contacts. Internally the OMG is composed of large sheets of Group 1 and 2 granodiorite and biotite granite which were been intruded by tonalite whilst they were still at high temperatures. This interpretation fundamentally disagrees with an earlier one by Crane (1984) which modelled the pluton as being intruded as a single body which differentiated to produce the internal
compositional variations.

(9) The emplacement mechanism, which is by large scale sheeting, suggests that the granitic magmas were hot mobile liquids during their intrusion. They may have contained low crystal percentages (probably epidotes and early plagioclases) which would imply that the magmas where emplaced below the critical melt percentage. It has been estimated that the main granodiorite would require approximately 320 000 years to pass through the critical melt percentage after which it should behave as solid crystalline granite.

8.3 Major conclusions to Chapter 4

(1) During the late magmatic and sub-solidus history of the OMG a series of structures were produced; these have been described as a chronological sequence which has been constructed from individual overprinting relationships throughout the pluton.

(2) A series of discrete sinistral shears were the first structure to form. These are narrow (1cm) shears which displace the tonalite and granodiorite/biotite granite contacts. They are overprinted by the main OMG foliation and are unusual in that they represent a localized shearing prior to the main pervasive ductile fabric development. They are generally formed in the sheeted contact zones and may have a sinistral displacement up to 4.8m. These shears are mainly extensional in geometry and may occur in linked systems which display strike-slip duplex characteristics such as branch points, tip zones, hangingwall and footwall ramps.

(3) The main foliation is strongly developed and is present in all the major components and most minor components. Its strike is approximately NE-SW with a dip between 48° NW and vertical. It is formed by a grain shape alignment of quartz, biotite, alkali feldspar and plagioclase. It is completely
penetrative and cross-cuts all internal lithological boundaries. There is an associated gently NE or SW plunging stretching lineation which is parallel to the long axis of dioritic xenoliths and is therefore represents the $1+e_1$ direction of the finite strain ellipsoid. The geometry of the foliation in all areas is a $S>L$ foliation. The foliation can be classified as a solid state foliation as all the mineral phases are deformed. There may be evidence for the pre-existence of a magmatic state foliation.

(4) The main foliation is deformed by a set of sinistral shear bands, which together with the main foliation, form the development of S-C fabrics extensively throughout the pluton. The shear bands often form in dextral and sinistral conjugate sets. The shear bands are extensional in geometry and consist of up to 1mm wide, elongate ribbons of quartz and feldspar. A total displacement of 3.4km has been estimated to be directly attributable to movement on the shear bands. Data suggest that the S-C angle decreases, the shear plane spacing decreases and the C plane length increases with increasing finite strain.

(5) A major fold has been mapped from the change in vergence between the discrete sinistral shears and the tonalite contacts. This fold is upright with an axial plane which is parallel to the main foliation. Mesoscopic folding is mainly formed in fine scale sheeted zones and in the late pegmatites and quartz veins. These folds are also generally upright and the main foliation is axial planar to them.

(6) Late sinistral shear zones form as a lower temperature localization feature of the main foliation and sinistral shear band formation. They produce thin zones of mylonite (up to 3cm wide) which form strike-slip duplexes on a small scale (20cm) or a larger (15m) scale. These may be extensional or demonstrate hangingwall contraction and oblique overthrust movement.

(7) There is also a series of late dextral shears which produced folding
of tonalite contacts and offset pegmatites. One examples suggests that they were formed synchronously with the late sinistral shear zones as a result of late stage partitioning of the ductile strain.

(8) Strain measurements using xenolith shape ratios suggest the existence of high strain zones within the OMG, however the pluton contact zones indicate a moderate finite strain which may discount a forceful emplacement model for the OMG. Attempts to estimate strain from the spatial distribution of phenocrysts was unsuccessful because of the high finite strains in the OMG and the manner in which the strain is partitioned into conjugate shear bands. Data presented show how the fabric intensity may vary with the finite strain.

(9) The major late fault is the Knockaskibbole fault system which may have originally developed as late ‘break back’ shearing of the main sinistral deformation in the SW part of the pluton. This was subsequently reactivated, perhaps more that once and probably in a brittle manner in the late Carboniferous.

(10) The chronological sequence which developed in the Ox Mountains is thought to provide good evidence for the continuation of sinistral ductile shearing as the pluton cooled with the deformation increasingly partitioned into localized high strain zones.

8.4 Major conclusions to Chapter 5

(1) Previous work has suggested that the the Lough Talt Adamellite and the Easkey Lough Adamellite are considerably younger than the Ox Mountains Granodiorite. These plutons have been dated at $401\pm33$Ma using Rb/Sr whole rock dating.

(2) The LTA is poorly exposed and the contact with the OMG has not been observed. The contact with the metasediments is however exposed and
is steep, NE-SW trending adjacent to the pluton. There is also evidence for a floor contact where the LTA is situated above a gently SW inclined contact with the metasediments underneath.

(3) The LTA contains a solid state grain shape alignment fabric. This is an intensely developed S > L fabric parallel to the steep contact, which decreases in intensity and its dip within a short distance from the contact. Above the floor contact is a zone of intense L fabric development suggesting that a type of strain superposition has taken place in this region.

(4) The LTA contains numerous dioritic xenoliths in an otherwise homogeneous pluton, which permit estimation of the finite strains. The strain profile shows a general decrease in X/Y ratios moving inwards from the contact zone which is interrupted by narrow high strain zones which correspond to localized sinistral shear zones in the LTA.

(5) A model for the LTA can be developed, in which a contact parallel foliation, perhaps related to pluton emplacement is modified by later sinistral shearing. The late shearing forms a high strain zone at the steep contact where the two fabrics are parallel and a L fabric above the floor contact where the fabrics are at a high angle.

(6) The ELA is a poorly exposed, apparently homogeneous pluton, however the contacts with the metasediments can be mapped in three areas.

(7) The solid state, NNE striking foliation in the ELA generally increases in intensity from the eastern side where it is a very weak, shallowly inclined fabric, to the western side where it dips steeply to the west. The presence of sinistral shear bands in this latter region suggest that the western margin has been deformed by sinistral shearing. This disagrees with the interpretation of Andrews (1984) who suggested that the ELA was emplaced and deformed in a dextral pull-apart.

(8) Both adamellite plutons are deformed by late dextral, semi-brittle
deformation concentrated in localized zones.

8.5 Major conclusions to Chapter 6

(1) The microstructure exhibited by the discrete sinistral shears is one in which plagioclase exhibits recrystallized, medium grained ribbons. This indicates that the plagioclase has either been fractured and annealed, or that it is deformed by a high temperature and pressure plastic deformation mechanism.

(2) The discrete sinistral shears may have formed by two processes.

(i) The fracture of a crystalline framework caused by pore fluid build up in the residual melt which is trapped in the framework.

(ii) A type of ductile instability may occur if the rate of change of shear stress with respect to strain is negative.

Plagioclase and alkali feldspar appear to have responded differently during the development of the OMG main foliation; this is seen in their varying microstructures. Plagioclase phenocrysts largely behave in a passive manner but may form large kinks which divide some of the crystals. They may also fracture if the stress build-up becomes too great. Alkali feldspar deforms more readily and is usually elongate. Evidence for crystal plastic deformation is provided by the observation of uneven extinction in most alkali feldspar crystals and by the development of myrmekite and recrystallized quartz grains on grain margins. Plagioclase microstructure in the LTA and the ELA display brittle fractures and do not show evidence that they have been deforming by crystal plastic deformation mechanisms.

(4) Quartz microstructure in the OMG during the main foliation shows differing degrees of dynamic recovery. Samples from the NE part of the pluton may have been deformed at lower temperatures and pressures than the rest of the pluton because quartz is poorly recovered. The granodiorite to the SW
displays a higher temperature microstructure with the widespread development of elongate quartz ribbons. Quartz microstructure in the adamellites displays deformation lamellae and elongate sub-grains which have not undergone dynamic recovery.

(5) Biotite displays a recrystallized microstructure in the OMG with the development of very elongate ribbons, whereas it is folded and sheared in the adamellites and sub-grains are formed.

(6) Amphibole in the Ox Mountains diorites displays evidence that it was dynamically recrystallized to form sub-grains, providing additional evidence of high temperature deformation.

(7) Deformation localization appears to occur to a greater extent in the granites and granodiorites than the tonalites and diorites. This is likely to be due to the presence of alkali feldspar in the former lithologies.

(8) The localization appears to take place by a process involving the production of myrmekite at grain margins perpendicular to the maximum stress orientation in the foliation. The myrmekite is produced by the build up of strain hardening at alkali feldspar megacrysts boundaries. This disturbs the equilibrium and a replacement of alkali feldspar by plagioclase and quartz occurs. This process involves a 10% decrease in volume which releases the high normal stress on the grain boundary and may indicate a change in deformation mechanisms from crystal plastic to diffusive mass transfer. Potassium ions released during this process may form muscovite which is developed along the C-planes in the granodiorites.

(9) Temperature estimates using the two feldspar geothermometer indicate that the localization formed at lower temperatures compared to the main foliation.

(10) Environmental conditions suggested for the main foliation and S-C fabrics indicate lower amphibolite facies deformation in the OMG. Deforma-
tion localization in this regime is controlled by a temperature and pressure de-
pendant switch of the predominant deformation mechanism in alkali feldspar. 
Higher temperature steady state flow with recovery gives way to strain hard-
ening which induces a diffusive mass transfer process. If this process is con-
tinued the entire granite texture becomes a fine grained mylonite with relict 
plagioclase crystals.

8.6 Major conclusions to chapter 7

(1) The emplacement of the Ox Mountains Granodiorite in a transcurrent 
shear zone, pull apart or releasing bend structure is discounted. The dyking 
or sheeting contact in either of these structures are expected to be steep and 
orientated at a high angle to the shear zone boundary. The original contact 
orientation is likely to have been originally gently to moderately to the NW 
and SE, with a strike orientation which is sub-parallel to the main shear zone 
transport direction.

(2) The brittle intrusion related structures in the contact zones 
have a movement sense which is up towards the centre of the pluton (and 
shear zone). The intrusion of magma in these structures creates a vertical 
extension direction and the $1 + e_2$ direction becomes sub-horizontal during 
the intrusion related deformation and is parallel to the intersection between 
the granodiorite sheets and the main wall rock fabric.

(3) The tonalite dykes may have been intruded as a conjugate set and 
their original orientation implies that either $\sigma_1$ or $\sigma_3$ is vertical, thereby im-
plying that $\sigma_2$ is horizontal. This may contrast with the stress system in a 
transpressional shear zone such as the Ox Mountains, where $\sigma_2$ is expected to 
be vertical.

(4) The two conclusions above indicate that during intrusion the central
part of the shear zone was an area of vertical extension and dilation. This region is bounded on both the northern and southern flanks by high strain zones. Oblique upward movement on these structures produces an area of vertical dilation in the core of the structure where the two high strain zones meet. The granitic magmas are likely to ascend shear zones in the central block which is descending relative to the shear zones on either side.

(5) The accession of the magma created an interruption in the ductile pervasive shearing which is known to have pre-dated the pluton emplacement and strongly deformed the pluton afterwards. This may have been due to the influence of the magma on a dilational regime which might have lower differential stresses than the normal transpressive shearing.

(6) The adamellites are likely to have been emplaced in an extensional cavity created by reactivated movements on the Glennawoo shear zone. Following uplift this shear zone is likely to have behaved in a less ductile manner than during its early history and may have created the extensional cavities by differential movement on semi-brittle detachments in the releasing bend sites.

(7) Granitoid rocks are more commonly associated with transpressional shear zones rather than transcurrent shear zones. Granites are also associated with oblique collisional zones and more orthogonal regions of crustal thickening.

(8) In general granites are produced by: fractionation of basic melts, crustal thickening and anatexis, hybridization and mixing of partial melts of underplated diorites in subduction zones and by adiabatic decompression associated with uplift.

(9) The Ox mountains Granodiorite magmas may have been produced by a crustal thickening event related to a perturbation of the thermal structure of the continental lithosphere by either the obduction of a major ophiolite slab or a major Caledonian thrusting event. The Ox Mountains shear zone may
have triggered the melting process in this perturbed region and then provided a conduit for the accession of localized batches of the magma.
REFERENCES


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THE OX MOUNTAINS GRANODIORITE LITHOLOGICAL AND STRUCTURAL MAP

STRUCTURAL SYMBOLS
- Main foliation
- Vertical foliation
- Metasediment fabric dip
- Contact dip
- Mapped contact
- Inferred contact
- Fault

KEY
- Group 1 Granodiorite
- Group 2 Granodiorite
- Brecciated Granodiorite
- Biotite Granite
- Muscovite Granite
- Tonalite
- Metasediments
- Sheeted tonalite and host rock

K. McCaffrey 1989
STRUCTURAL MAP OF THE CENTRAL & SOUTHWEST OX MOUNTAINS, W. IRELAND.

STRUCTURAL SYMBOLS
- F2 - L2
- S2
- D2 isochastic contact
- F2 with vergence - L3
- S3
- Overthrust sense
- Lough Tall Side
- Glenawoo Side
- North Ox Mountains Side
- Cailow Shear Zone
- Tawneyshane High Strain Zone
- Oblique drop zone (Lough Easky Side)
- F4
- F4
- Shewater Thrust
- Fault
- Contact (exposed)
- Contact (interbedded)
- Uncertainly
- River
- Main Road

KEY
- Clonongovan Formation
- Upper Lisморan Formation
- Cailow Member
- Lower Lismoran Formation
- Ummoon Formation (includes the Newantrim Member)
- Leckee Transition Member
- Leckee Quartzitic Formation
- Tawneyshane Marble Member
- Tawneyshane Pumice Member
- Pre-caledonian basement
- North Mayo Inlier Dalradian
- Austroite
- granodiorite
- tonalite
- granite
- marble
- phosphate
- serpentinite