Multiple sheeting as a mechanism of pluton construction: the main Donegal granite, NW. Ireland

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MULTIPLE SHEETING AS A MECHANISM OF PLUTON CONSTRUCTION: THE MAIN DONEGAL GRANITE, NW. IRELAND.

By

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Declaration

No part of this thesis has been previously submitted for a degree at this university. The work described in this thesis is entirely that of the author, except where reference is made to previous published or unpublished work.

Alun R. Price,
Univeristy of Durham,
Department of Geological Sciences,
August 1997.

To

Mum & Dad

Frontispiece

"On a hillside in Donegal: a glimpse into the great Earth Cauldron"

Grenville Cole (1906)

A view of the "Earth Cauldron" from the summit of Moylenanav looking northwards into the Sruhanavarnis Valley, where the present author "spent many a summer night" during the summers of 1994-5

Abstract

This study is a detailed investigation concerning the construction of granite plutons by the incremental emplacement of granitic sheets. The modern consensus is that sheeted plutons are often controlled by tectonic structures such shear zones. The Main Donegal Granite (MDG), NW Ireland forms the basis to this study. This pluton is the largest presently exposed member of the Caledonian Donegal Batholith (~405 Ma). Field evidence from this highly deformed pluton, attest to emplacement along the long-axis of a sinistral transcurrent shear zone. The presence of long and persistent xenolith "trains" within the pluton has been taken as evidence of an overall sheeted structure; however detailed maps have not been available to test this hypothesis.

Two earlier members of the Donegal Batholith, the Ardara and Thorr plutons, whilst having their main outcrops outside the MDG, also occur as xenoliths within the main body. It can be demonstrated in a number of critical situations that these xenoliths are commonly more deformed than the host MDG facies. Furthermore the presence of original country rock contacts implies these xenoliths were originally in situ. These features imply that the shear zone was active prior to the emplacement of the MDG, with it controlling the emplacement of substantial parts of these earlier plutons. Further evidence from the study of parts of the petrographically similar and younger Trawenagh Bay Granite implies the sinistral shear zone was still operational after the majority of the MDG had crystallised.

New, detailed (scale 1:250) and reconnaissance mapping of the MDG, reveals its hitherto unrecognised heterogeneity. At least seven major plutonic zones or packages have been identified. All these units have an NE -SW elongate form parallel to the long axis of the pluton and are often, but not always, separated by extensive "raft-trains" of country rock and older plutons. The major packages in the central regions of the pluton are often complex and are composed of three main granitoid phases, ranging in composition form early granodiorites and tonalites to latest porphyritic and to lesser extent equigranular, monzogranites. The early granodiorite and tonalite sheets are now only preserved as xenolithic rafts within the later monzogranites. The broad range in composition/chemistry together allied with field observations implies a complex intrusion history, with these granitoid packages representing sites of long-standing intrusion within the pluton. In contrast, towards the more marginal areas of the pluton there are large units of monzogranite which are characterised by general homogeneity, but in reality are believed to consist of relatively small compostionally similar sheets. On all scales, either meta-sediments, older plutonic material, or early MDG facies are found to lie along the boundaries of younger intrusive units. This implies the pluton is primarily sheeted in character and that the "raft-trains" are partially disrupted, in situ roof material which has been wedged apart during the intrusion of the sheets.

The appearance of sheets within the field is dependent on the rheology of the material into which the granitic material was intruded into, i.e. to what extent has the host was crystallised. The degree of crystallisation in the host is related to how fast later sheets were being intruded, i.e. the rate of emplacement. The field relationships, in the central regions of the pluton, between the granodiorites-tonalites and the later monzogranites, are interpreted as representing zones of episodic-to slow emplacement, where earlier phases had become essentially competent by the time later units were intruded (i.e. capable of fracture). These earlier phases may be preserved as angular rafts within later sheets. At moderate emplacement rates earlier sheets may still be crystallising but sufficiently viscous to prevent mixing, except at their immediate boundaries with transitional contacts developing. The more homogeneous zones are believed to be related to rapid emplacement with original contacts betweeen pulses being destroyed at the level of emplacement due to homogenisation of pulses which had similar viscosities and hence allowed mixing.

The emplacement of granitic melts within active shear zones can lead to the development of a self-perpetuating situation, where melts in a shear zone will enhance deformation rates and cause greater displacements subsequently allowing more melt to enter the shear zone promoting even greater displacement rates. This process is only halted when melts within the source regions are drained; hence the rate of pluton construction and appearance of sheets within plutons is ultimately related to how fast granitic melts are being generated within the source regions.

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Chapter 1

Granite Ascent, Emplacement and Deformation: an overview

1:1 Introduction

It is well known that most granitic rocks are associated with zones of tectonic activity, most notably along continental margins. Granitoids commonly occur within orogenic regions involving continental crust and these include subduction along continental margins, continent-continent collision, zones of extension and transcurrent shearing. To a lesser extent granitoid rocks are found in association with ocean plate settings such as island arcs (diorites) and also the volumetrically minor plagiogranites found within ophiolitic complexes (Pitcher 1993). Generally granites will form in considerable volumes where the crust has been heated by a hot mantle source or where the continental crust has been thickened by collision processes (Fyfe 1988).

Subduction along continental margins produces the greatest volume of granite material (tonalite and granodiorite) with the western Cordilleras of North and South America being the most spectacular examples, e.g. the Peninsular Ranges Batholith of Baja California (Silver & Chappell 1988) and the Coastal Batholith of Peru (Pitcher 1978; 1979). The continent-continent collision zones tend to produce the "true" granites with the Himalayan Leucogranites and the Hercynian granites of SW England being typical examples. There are the post collision granites (*sensu stricto)* and granodiorites which tend to localise along intercontinental transcurrent shear zones (including transtension and transpression) e.g. the Caledonian granites of the British Isles (Hutton (1982); (1988); Hutton & Reavy (1992); McCaffrey (1992) and Jacques & Reavy (1994) D'Lemos *et al.* 1992)). Granites emplaced within intra-cratonic areas are associated with extension with subsequent mantle updoming acting as heat source. Granite types in these environments commonly have, but not always, peralkaline affinities, e.g. the Sara-Fier Granite Complex, Nigeria and the Tertiary centred complexes of the British Isles (Pitcher 1993).

From the distribution of the different granite types it appears obvious that the ambient tectonic regime is going to strongly influence the generative processes and nature of the source materials (Pitcher 1993).

1:1:1 Petrogenesis of granitoid rocks

It was the work of Chappell & White (1974; 1984 & 1992) on the Caledonian granites of the Lachlan region of SE Australia that shed light on the source of granite rocks based on the chemistry. These authors recognised two main granite types (summary taken from Pitcher 1993). The "S" type granites which tend to be more chemically evolved and generally potassium-rich are believed to have been derived from crustal material that has been through the erosional-sedimentary cycle. The "I" type granites are chemically more primitive and contain generally less potassium and are derived from igneous precursors that have not been previously recycled. The identification of these source rock signatures therefore provides a tool in discriminating granite types in different tectonic environments, but as Pitcher (1993) strongly emphasises this discrimination should never be used in isolation from petrology and geological criteria. The use of isotopes Sr_i ($^{87}Sr/^{86}Sr$) and ε_{Nd} ($^{143}Nd/^{144}Nd$) signatures enables the nature of the source for granitic rocks to be identified. The presence of even more geochemically primitive, potassium poor calc-alkaline rocks than the "I" types of Chappell & White (1974) within island arc settings has led to the labelling of such granitoids as "M" types. Such granites have almost exclusive mantle affinities. Therefore the "S" and "M" type granites would appear to be end-members with the "I" types intermediate between these end-members (Pitcher 1993). Based on the fact that tectonic setting and source are interrelated then it may be possible to broadly correlate composition with tectonic setting. The "M" type granites are typical of island arcs (quartz diorites and tonalites), whilst the plate margin granodiorites and tonalites will be mainly "I" types with M-I transitions. The "S" types peraluminous granites are typical of tectonic settings during the early stages of continent collision. The post orogenic uplift granites show characteristics of both "S" and "I" granites. In contrast, the alkali granites of the intra-plate anorogenic environments, despite having a dominantly mantle source are grouped as "A" type granites.

1:1:2 Emplacement:- over the past two hundred years or so there has been extensive work carried out on attempting to find a solution to the so-called "granite problem". Were granites igneous, metamorphic or metasomatic in origin? The "ins and outs" of this problem is beyond the scopes of this thesis (see Pitcher 1993 for historical perspective) although the investigation of this problem has promoted much of the current understanding on ascent and emplacement mechanisms (England 1988). Evidence for igneous origin was obtained from field studies where granite bodies were clearly seen to be intrusive, often having discordant contacts with their host country-rocks, (Cloos 1923; Balk 1937). The major problem with intrusive granites was "the space problem" where the volume of the intrusion was larger than what could be explained by structures in the country rocks. For this reason there were opponents of magmatic origin, favouring metamorphic or metasomatic processes which changed the rock in-situ and therefore did not constitute a space problem. Sederholm (1907) favoured this opinion and introduced terms such as migmatite, anatexis and palingenesis to the literature to describe rocks formed by partial melting of crustal rocks (Pitcher 1993). It was obvious that the views of individual workers were based on their own field observations in variety of settings which are now known as orogenic belts (England 1988). Eskola (1932) believed all granites to be magmatic, and classified them as syn-, late- or post tectonic based on their association with phases of regional metamorphism and deformation. Syn-tectonic granites were gneissose and concordant to country-rocks (often gradational) and contained basic inclusions; late tectonic granites were discordant (or slightly concordant) and rarely gneissose, whilst post-tectonic granites were unfoliated and intruded after deformation has ceased. This classification was extended by Read (1957) with the "Granite Series" and by Buddington (1959) who implied the genetic relationship between granite type and crustal level. With the "Granite Series" of Read (1957), the migmatites (autochthonous granite) were formed at the deepest levels of orogenic belts during peak tectonic activity. Discordant and partially discordant foliated granites (parautochthonous) were intruded at higher levels during later stages of orogeny and were believed to have separated from the migmatite complexes at depth. The unfoliated, discordant granites (magmatic-intrusive) were emplaced at high levels of the crust after deformation had ceased. Buddington (1959) extended the "Granite Series" but rejected the time dependent nature of classification by Read (1957). Buddington (1959) classifies granites at three levels of emplacement zones within the crust. The autochthonous granites of Read (1957) were reserved to the "catazone", parauthochthonous granites were assigned to the mesozone, whilst the magmatic intrusive granites and plutons were classified in the epizone. The three zones were depth related but Buddington (1959) believed they were intruded at all levels during orogenesis.

The work of Pitcher & Berger (1972) on the Donegal Batholith showed that granites of the same age, emplaced at the same level in the crust showed differing styles of emplacement, a feature incompatible with the "Granite series" approach. Further work by Pitcher (1978; 1979) showed that granite type and emplacement style may well be related to the ambient tectonic regime. The new tectonic approach to granite emplacement was applied by Hutton (1982) on the Main Donegal Granite and extended to other tectonic situations in later papers (Hutton 1988). This author implied that "..*in general, space for magma can be created by a combination of tectonically created cavities and internal related magma buoyancy*".

Granite plutons display a wide variety in size, shape and structural diversity. The structures which one sees within a pluton are more likely to reflect the emplacement mechanisms rather than evidence of the ascent mechanisms from the source. Such emplacement structures preserve the interactions of the wall-rock and the cooling granite in response to internal buoyancy forces and external tectonic forces (Hutton 1988). The way in which such structures manifest themselves within granitic rocks will be governed by the physical properties of the magma. The physical properties or rheology of granitic magma shows marked changes in viscosity during its crystallisation and therefore ones needs to understand the varying structures which develop in response to this increase as the granitic pluton cools. The following sections will discuss such features.

1:2:- The Rheology of Granitoid Magmas

The science of rheology studies the flow of matter in response to applied stress and is thus concerned with material properties, strain, strain rate and environmental factors. It has been shown that materials display three main types of behaviour when subjected to applied stress.

1) *Elasticity*:- this occurs when stress (σ) is proportional to strain (e), where after the stress has been removed the material recovers and does not record any evidence of strain. On a graph this plots as a straight line which obeys Hooke's Law, where,

$$\sigma = \mathbf{Ee}$$

The E value is Young's modulus and is the gradient of the straight line and reflects natural variations in resistance of different rock types to elastic deformation, i.e. Young's Modulus is a measure of "stiffness" within a rock.

2) *Yield (or plastic behaviour)*:- above a certain value, the yield strength (or elastic limit), a material no longer behaves in elastic fashion with it recording permanent strain, i.e. it deforms. Whether or not the material deforms in a brittle or ductile fashion is controlled by temperature, confining pressure and strain rate, factors which also effect the yield strength value. At very slow strain rates, materials (time dependant for different materials) may deform permanently below the yield strength by the process of "creep" (Davis 1984) (see section 1:3:3:1) on deformation mechanisms). In granitic rocks over long geological periods "creep" is an important

process. Therefore the yield stress is a transgression from deformation being non-permanent to permanent (McErlean 1993).

3) *Viscous Flow*:-viscous materials can be imagined as liquids that flow when subjected to any differential stress. The viscosity of a material is a measure of its resistance to flow. Unlike plastic materials truly viscous materials have no overall fundamental strength threshold which would need to be overcome before flow can occur. In an ideal (Newtonian) viscous material the relationship between stress and strain rate is linear where

$$\sigma = \eta\varepsilon$$

(σ = differential stress, η= viscosity constant and ε= strain rate)

The above equation is obeyed by materials which show Newtonian flow and these include most pure fluids (e.g. water). It is known that most upper crustal rocks do not behave as Newtonian fluids when a stress is applied although magmas have been modelled in this way especially, where melts contain crystal percentages of <35-50% (Shaw 1965; Holtz et al 1996). In fact it was Scrope (1872) who coined the term "magma" because it was apparent that it contained a mixture of liquid and solid particles.

Materials can display more complex rheologies by a combination of the above mentioned behaviours (in Twiss & Moores 1992) e.g. visco-elastic:- permanent strain accumulates viscously and begins as soon as stress is applied. Highly viscous materials will behave elastically for loads of short duration but like a viscous material for long term loads, e.g. silly putty. Elastic-plastic materials show a combination of elastic and permanent strain, but before plastic deformation occurs the yield strength must be exceeded. Release of stress allows elastic recovery although permanent strain remains. Visco-plastic (Bingham) materials (e.g. blood and wet paint) only display linear viscous behaviour after the yield stress for plastic materials has been achieved. The rapid application of relatively high stresses may cause a material to fracture where stress is then released and deformation may continue in a plastic fashion (Davis 1984). Pitcher (1987), Pitcher & Read (1960a) showed that granites may behave in this fashion as seen by the intrusion of syn-plutonic dykes which become deformed by subsequent movement in its host. Figure 1:1 shows the behaviour of these various rheologies when shear stress is applied.

Granitoids display a wide range of physical properties throughout the duration of their crystallisation history. Viscosity is the main property which affects the rheology of granitic magmas and this itself is controlled by temperature, pressure, composition, crystal content and the presence of volatiles, most notably water (Shaw 1965). In the source region of granitic melts it is the first three factors which primarily control magma viscosity (Petford 1996), whilst at the emplacement level the

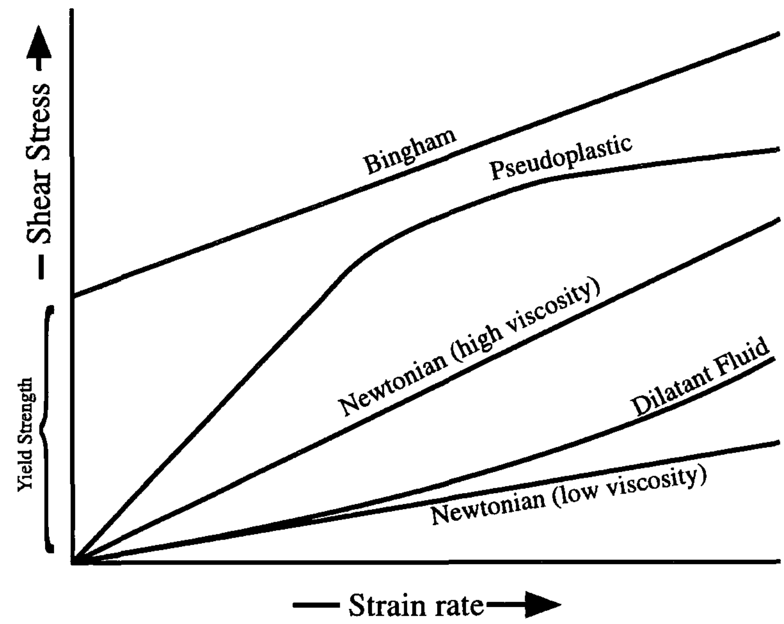


Figure 1:1
Graph showing shear stress v strain rate for varying rheologies (modified from Ingram 1992)

fraction of crystals becomes more important to overall magma viscosity. At crystal percentages of <35 % the effective viscosities are generally similar to crystal free melts and can be essentially modelled as Newtonian fluids, whilst at crystal contents between 35-55% there is only a gradual increase in apparent viscosity (Petford 1996) (generally less than one order of magnitude (Shaw 1965). At crystal contents greater than 50-55%, depending on the composition and presence of volatiles, which control viscosity of the melt fraction and the size and shape of the suspended crystals, there is a rapid increase in the viscosity (many orders of magnitude) of the magma, with it no longer obeying a linear relationship with the material starting to show Bingham body-pseudoplastic behaviour (Arzi 1978, van der Molen & Paterson 1979; Shaw 1965; Holtz *et al.* 1996)). This increase in viscosity is related to the increasing interaction of the suspended crystals until eventually "lock-up" occurs with the strain being transmitted along crystal contacts with the solid phases starting to deform in a plastic fashion. This transition is termed the Rheological Critical Melt Percentage, (RCMP) by Arzi (1978) which for most rock types is between 20 ±10% melt fraction, although a value of 30-35 % was obtained from experimental partial melting of granites by van der Molen & Paterson (1978) (see figure 1:2). Above the RCMP the solid framework will deform in a plastic manner, e.g. dislocation creep (Tullis and Yund 1980). From the above authors (Arzi (1978); van der Molen & Paterson (1979)) it was the belief that there was a unique RCMP which existed at similar values for the liquid-solid transition (i.e. crystallisation) and for solid-liquid transition (i.e. melting). Vigneresse *et al.* (1996) consider the rheology of a granite magma from the onset of partial melting at the source to the crystallisation at the site of emplacement with the solid-liquid and liquid-solid transitions being rheologically

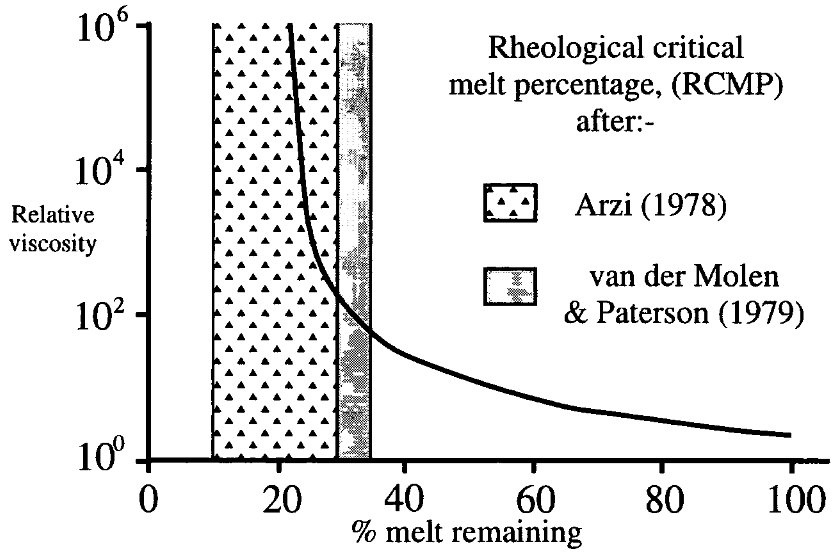


Figure 1:2 Curve showing the increase in relative viscosity as crystal percentage increase. Shading represents the rheological critical melt percentage (Arzi 1978; van der Molen & Paterson 1979).

different, hence disproving the existence of a unique RCMP. The above authors suggest four processes occur (taken from Vigneresse *et al.* (1996)):-

i) Liquid percolation threshold (LPT):- to overcome this threshold there needs to a minimum of 8% volume melt. Above this value melt pockets can connect allowing magma displacement and very localised migration, e.g. partially molten domains within migmatites.

ii) Melt escape threshold (MET):- above 20-25 % melt segregation and escape can occur which allows transport over large distances i.e. source to level of emplacement.

iii) Rigid percolation threshold (RPT):- this occurs during emplacement and crystallisation. Within a magma with < 20% solids particles can freely rotate independent of each other to define a fabric. Between this value and ~ 55% solids they progressively interact to produce a rigid skeleton. This upper value is the RPT beyond which the particles sustain stress, although the liquid fraction can still flow. Rearranging of these particles is by the localised development of incipient shear zones.

iv) Particle locking threshold (PLT):- this occurs when the system totally locks up when random close packing is reached which is believed to be at 72-75 % solidification.

Therefore the first two processes apply to melting whilst the latter two apply to crystallisation (Solid → Liquid → Solid). It is clear that the two transitions are not the reverse of each other. The values of LPT and RPT are independent of external forces but relies on abundance and type of minerals forming the matrix in which melt connectivity is developing (Vigeneresse *et al.* 1996). These same authors state the values of the MET and PLT will be controlled by external forces, such as deformation, and also by the shape of the particles.

The existence of the RCMP during partial melting of granitic compositions has been doubted by experimental work by Rutter & Neuman (1995). During melting experiments on fluid-absent Westerly Granite, these authors found no evidence of a change in rheological behaviour (i.e. sudden decrease in bulk strength), as the RCMP predicts, corresponding to the break-up of the solid framework into the increasing viscous melt where the latter then controlled the overall mechanical behaviour of the aggregate. Rutter & Neuman (1995) stated that the RCMP may exist in systems where the melt phase has a low viscosity.

Within the realms of this thesis the behaviour of granite melts is mainly concerned with the structures that form during the crystallisation of a granitic melt. In this thesis the terminology of Vigneresse *et al.* (1996) will be used in appliance with the RCMP phenomena due to its greater relevance when concerned with field analysis of structures within granitic rocks.

The behaviour of granitic melts is significantly different to that of basalt melts, mainly because of their viscosity, with basalts (10^1-10^2 Pa s) being 3-6 orders of magnitude less viscous than granitic melts. Within the granitoid field itself there are considerable variations in viscosities ranging from 10^4-10^8 Pa s, when measured as crystal free melts, (Petford 1996; Baker 1996). The tonalites and granodiorites of the Cordilleran batholiths are generally the least viscous with the leucogranites being the most viscous whilst syn-orogenic granites have intermediate viscosities (Petford 1996).

The above range in viscosities for granitoid rocks will therefore strongly affect the structures which develop within granitc melts during emplacement and as they cool to the ambient temperature of their countryrocks.

1:3 The Formation of Granitic Structures

The need to identify and study structures within a granite pluton and its wall rocks are vital in trying to establish the tectonic processes that were occurring during emplacement and throughout the crystallisation history and therefore it is necessary to identify structures that developed at different stages of the cooling history. These can be conveniently divided into magmatic, sub-magmatic and solid state features.

1:3:1 Magmatic state Stuctures - PFC fabrics

Magmatic fabrics are produced when the percentage of crystals to viscous melt within a granitic magma is relatively low, (usually below the RCMP on the Arzi curve). Commonly the crystal suspension (± enclaves of either country-rock or older igneous material) will be composed of early euhedral plagioclase feldspar phenocrysts and to a lesser extent hornblende or biotite. These inclusions will be rotated (the degree of which depends on the axial ratio) towards some type of alignment governed by the ambient stress field. If deformation ceases at this point then the remainder of the melt will crystallise and will be undeformed. Therefore the fabric will consist of aligned feldspar and early mafic minerals which are internally undeformed, surrounded by an essentially unaligned and undeformed matrix (figure 1:3c) (Hutton 1988a). This type of fabric was termed a Pre-Full Crystallisation fabric (PFC fabric) by Hutton (1988a) and is equivalent to "magmatic state deformation" of Blumenfeld and Bouchez (1988) and "magmatic flow" by Paterson *et al* (1992). This fabric is distinguished from those structures formed above the RCMP by the general absence of intracrystalline deformation within quartz, the mineral most sensitive to strain when stress is applied.

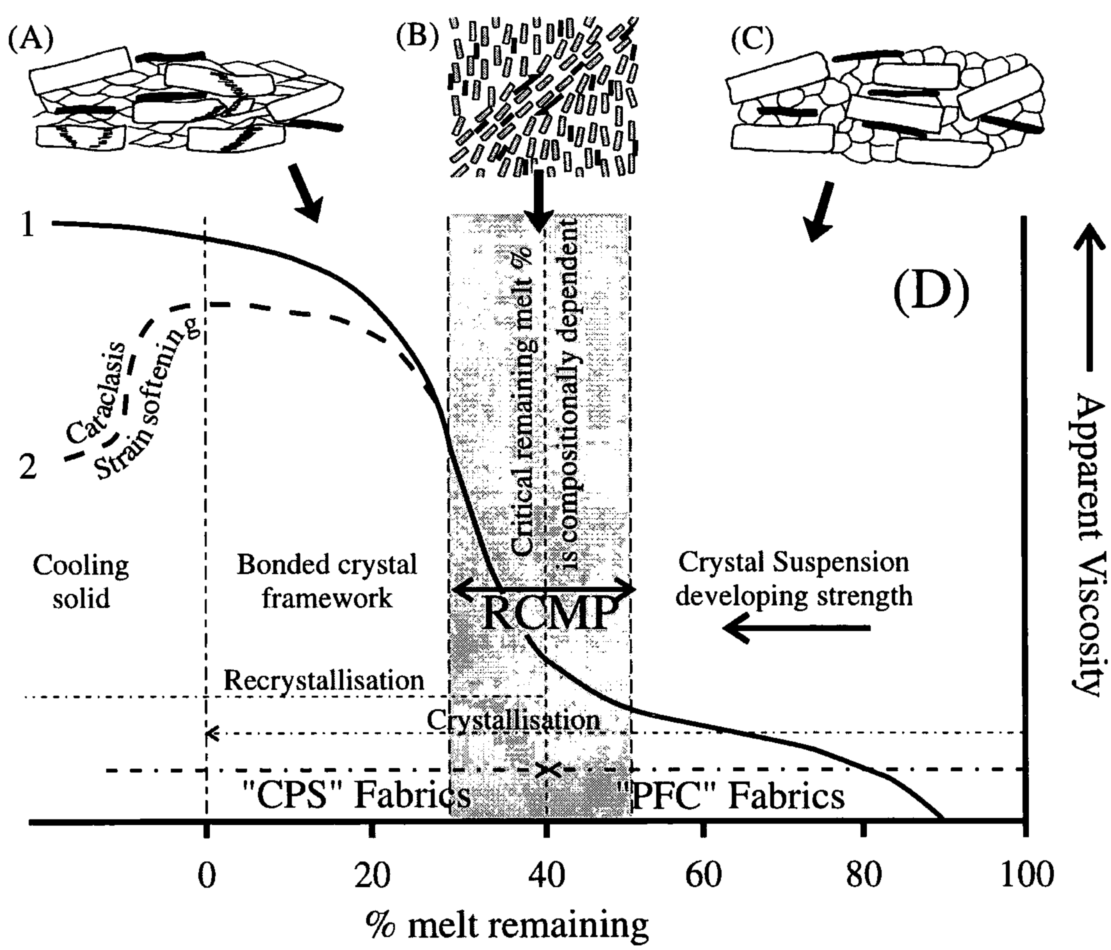


Figure 1.3 Granite fabric development during crystallisation; (a) CPS fabric; (b) PFC "lock-up" shears; (c) PFC fabrics; (d) Arzi-type diagram showing the structures which develop throughout the cooling history of granitoid magmas (modified from Pitcher 1993); (Hutton 1988; Ingram & Hutton 1994).

Other magmatic structures within granitoids include preferred alignment of microgranitoid enclaves, which are interpreted as incompletely solidified globules, that lack evidence of plastic deformation or recrystallisation (Paterson *et al.* 1989). Further evidence of magmatic deformation is the deflection of magmatic fabrics around meta-sedimentary encalves. This type of fabric would expect to develop in areas where a significant viscosity contrast exist between melt and solid phases (Berger 1971; Hutton 1982).

1:3:2 Magmatic state-solid state transition

When the crystal percentage approaches, or is at, the RCMP the suspended mush will start to behave as a solid framework. Hutton & Ingram (1992) observed that at this point just before the onset of plastic flow behaviour deformation is concentrated within the magmatic state into discrete shears which they termed PFC "lock-up shears". These appear as zones where undeformed plagioclase laths have been rotated in to the shear plane (see figure 1:3b).

1:3:3 Solid-state deformation of granitic rocks

Solid-state fabrics are those which form when the granite has cooled below the solidus of all respective mineral phases and may continue throughout the cooling period of a granitic pluton. Before going on to describe solid-state fabrics it is important to describe how granitic rocks deform with the following sections referring to behaviour of different mineral species at the grain-scale (i.e. thin section).

1:3:3:1 Deformation Mechanisms

The way in which granites accommodate strain is governed by a series of factors relating to environmental and lithological variables. These include the following:- mineralogy, grain size, presence or absence of fluid, temperature, pressure and strain rate (Bailey (1997). Combinations of these factors will govern the way in which a granite will deform and by what processes this will occur.

The various mechanisms by which rocks deform can divided into three main groups and relate to the variables mentioned above:-

i) Cataclastic flow (frictional grain boundary sliding, fracture and cataclasis)
ii) Diffusive Mass Transfer (pressure solution creep and diffusion creep)
iii) Crystal plastic processes (dislocation creep)

1:3:3:1a Cataclastic flow

The process which causes the creation of new surfaces and loss of cohesion due to fracturing (Knipe 1989).

i) Fracturing:- involves the development and propagation of cracks due to the failure of a rock after the elastic limit has been exceeded. The fractures then act as weaknesses which may then accommodate further displacement during ongoing deformation.

ii) Frictional Grain Boundary Sliding:- the sliding of grains past one another. This occurs where the coefficient of friction is overcome and cohesion between grains is lost. This process is enhanced in conditions of low confining pressure and high fluid pressure (low effective stress).

iii) Cataclasis:- involves the mechanical fragmentation of grains by microfracturing and subsequent frictional sliding, rigid body rotation and overall decrease in grain size with increasing strain. Due to this process being frictional (dependent on normal stress, σ_n) and dilatant, this deformation mechanism is pressure sensitive, i.e. confining pressure and fluid pressure (Stewart 1997).

1:3:3:1b Diffusive Mass Transfer (DMT)

This induces deformation by the transfer of material away from zones of high intergranular normal stress to zones of low normal stress, i.e. a "sink" (Knipe 1989). The driving forces for DMT processes are caused by the development of the following:- chemical potential gradients due to stress variations in the rock; fluid pressure gradients or variations in internal strain energies of the grains (Knipe 1989). Finer grain sizes are favoured for DMT processes as the diffusion path lengths are low and the differential stress is also low enough to inhibit crystal plastic deformation mechanisms (Knipe 1989). There are three main DMT mechanisms:-

A) Nabarro-Herring Creep:- the use of the crystal lattice as a diffusive medium by the migration of vacancies through the lattice.

B) Coble Creep:- the use of grain boundaries as the diffusive medium.

Both A) and B) are commonly grouped under **diffusion creep** and tends to predominantly occur at high temperature regimes.

C) Pressure Solution Creep:- diffusion along grain boundaries where a film of aqueous fluid is present. Highly strained parts of the lattice, i.e. grain-grain contacts are more soluble causing material to migrate away from zones of high normal stress to be precipitated in zones of lower normal stress e.g. strain shadows. Pressure solution tends to occur at relatively low temperatures.

1:3:3:1c Intracrystalline deformation

There are two main processes which cause internal deformation within a grain; these are crystal plastic processes and deformation twinning.

Deformation (Mechanical) Twinning:- twinning is a common feature in plagioclase and also calcite but only occurs in certain crystallographic orientations and overall can only accommodate limited strain. Deformation twinning tends to occur in the lower temperature range of deformation and can be distinguished from growth twinning by variable twin thicknesses and tapering appearance (Passchier & Trouw 1996).

Crystal plastic deformation

This term has often been used to describe deformation resulting from both diffusion creep and dislocation creep as in some natural rocks it is sometimes difficult to distinguish between the two processes (Passchier & Trouw 1996). In this thesis the term crystal plastic deformation will be used to address structures formed mainly by dislocation creep. The following descriptions of crystal plastic processes are taken from Passchier & Trouw 1996).

Crystals usually have lattice defects which can be broadly divided into two groups i) Point defects:- these are either lattice vacancies or extra lattice points (i.e. interstitial vacancies) ii) Line defects:- these may be caused by extra half lattice planes within a crystal. The ends of such planes are known as "edge" dislocations. "Screw" dislocations from when one part of the of the crystal is displaced over one lattice distance causing the crystal to be twisted. For permanent strain to occur within a crystal there must be a relative change in position of the atoms or molecules within the lattice. Changes in position are accommodated by movements along lattice defects (dislocations) and leads to intracrystalline deformation. Ductile deformation of rocks due to the migrations of dislocations and vacancies within minerals is a very important and common process. Intracrystalline deformation involving movement along dislocations is called ***dislocation glide***. In thin section undulose extinction within a mineral is a characteristic feature of dislocation glide. Within different minerals dislocations have a distinct orientation in respects to the crystallographic axes and can only move on specific crystallographic plane and direction. A specific plane coupled with a slip direction is known as a *slip system*. Most common rock forming minerals have several slip systems which may be active at the same time. Changing conditions such as temperature and magnitude of the stress field may cause new slip systems to become active. Where several slip systems are operating this ultimately leads to migrating dislocations becoming entangled which may restrict further movement. Furthermore these tangles may block newly formed dislocations which may then pile up behind the blocked ones. Therefore the rock becomes harder to deform with this process being known as ***strain hardening***. Strain hardening tends to enhance brittle deformation as higher stresses are needed to deform the grain. At higher temperatures (0.5 homologous temp.) there are processes which counteract the effects of strain hardening by reducing the internal strain energy within a mineral and hence allows continued ductile deformation to occur within a mineral. An important mechanism is ***dislocation climb*** which allows dislocations to pass obstructions resulting from the migration of vacancies to dislocation lines. This effectively displaces the dislocation allowing it to climb over the blocked site. The processes of dislocation glide and dislocation climb are part of a combined mechanism known as ***dislocation creep***. Dislocation creep becomes more efficient at higher temperatures and lower strain rates. Dislocation climb becomes more dominant as homologous temperatures increase and allows more orderly configurations of dislocations to develop. Features indicative of dislocations forming lower energy arrays are deformation lamellae which are usually parallel arrays of lamellae with different refractive indices to the surrounding crystal. They correspond to zones of dislocation tangles and subgrains.

Recovery

This is the name given to the process which reduces the internal strain energy within a crystal. This is achieved by decreasing the dislocation density by reordering of dislocations into low energy arrays. This process is thermally activated as higher temperatures promote greater ordering of defects with dislocations tending to migrate to specific areas such as deformation bands or dislocation walls separating slightly misoriented areas of a crystal lattice. This progressive misorientation leads to the development of subgrains within a crystal (clearly visible in thin section, most notably in quartz). Subgrains have lattice orientations slightly misoriented from their neighbours (usually < 5%). Subgrains tend to have low internal energies as the dislocations have migrated to the boundaries which now form the subgrain walls. Hence these grains may appear less deformed, although they may well become subjected to further dislocation creep.

Dynamic Recrystallisation

This is the formation of new grains (neoblasts) during syntectonic deformation. There are two main types of dynamic recrystallisation.

- Grain Boundary Migration (GBM) Recrystallisation
- Subgrain Rotation (SR) Recrystallisation

GBM recrystallisation involves the migration of grain boundaries from a crystal with lower dislocation density into a mineral with a higher dislocation density. With increased strain the boundary migrates into the more deformed grain leaving a reordered, relatively strain free crystal, behind. This process is due to atoms or molecules being more readily accommodated in a more ordered lattice than within a more disordered lattice. Structures attributable to GBM in thin section are highly serrate or lobate boundaries. Occasionally where a less deformed grain has bulged in to a more deformed grain, the neck of the bulge may be pinched out to leave a new grain enclosed within an old grain.

SR recrystallisation results from the increasing accumulation of dislocations along a subgrain boundary during recovery. This leads to a progressive misorientation of the subgrain lattice. With increased rotation (>10°) the sub-grain boundary becomes a high angle boundary of a newly recrystallised grain. SR recrystallisation does not decrease the dislocation density with the newly formed grains and hence still show strong undulose extinction. SR recrystallisation is responsible for grain size reduction within a deforming rock. SR recrystallisation structures can be recognised by domains of relic deformed grains, subgrains and new grains which generally have similar lattice orientations, a feature identified by the use of a sensitive tint plate (Stewart 1997). During progressive deformation SR recrystallisation is thought to be more dominant early on and causes grain size reduction within a rock. As dislocation densities increase within the subgrains this initiates GBM recrystallisation with this reducing the overall strain energy within the rock minerals and hence obliterating many structures formed by SR processes (Passchier & Trouw 1996).

Static Recrystallisation

When deformation ceases within a rock the deformed minerals are generally not in a state of minimum internal energy despite recovery and recrystallisation processes having occurred. The crystals present will still contain dislocations and subgrains etc. If the temperature was not high enough then such fabrics will be preserved and remain unaltered during uplift and erosion. On the other hand if the temperatures were still sufficiently high enough (or fluid is present) after deformation has ceased then recovery and recrystallisation processes may still occur with internal strain energies being minimised and hence the mineral will appear strain free. Lattice defects are not the only contributor to high internal energies within a volume of rock; grain boundaries also have the same effect. Therefore a decrease in grain boundary area and increase in grain size, often at the expense of smaller grains, can also lower the internal strain energy. This leads to the development of polygonal crystal mosaics which have the most "strain free" configurations. In thin section these are easily identified with 120° triple junctions developed between crystals. This process is known as Grain Boundary Area Reduction (GBAR) and it is a process which occurs during deformation although its effects are most profound during static recrystallisation. In this study static recrystallisation has only occurred to a limited degree within the granites.

1:3:3:1d Mineral strength and deformation of polymineralic rocks

The above processes occur within almost all of the rock forming minerals depending up on the prevailing physical conditions. Due to different crystallographic structures and composition different minerals will behave in different ways. In respect to granites, the behaviour of quartz and feldspar is very important, as these minerals commonly comprise 80-95% of the rock.

Deformation of quartz-feldspar rocks: granites

The behaviour of rocks containing quartz and feldspar is highly variable and is strongly dependent on metamorphic grade. In figure 1:4 the behaviour of the main minerals within granitic rocks are considered. At very low grades (see figure for temperature ranges) both quartz and feldspar deform in a brittle fashion by microfracturing with quartz tending to be stronger than feldspar. The presence of cleavage within feldspar is believed to be the reason for its greater weakness in low grade conditions (Passchier & Trouw 1996). Under low grade conditions quartz is now deforming in ductile fashion by dislocation creep whilst feldspar is still behaving in brittle fashion. The strength contrast is now reversed with feldspar being mechanically more stronger under these conditions. Around feldspars "core and mantle" structures may develop with the cores tending to be fractured whilst the mantles are composed of fine grained feldspar resulting from either marginal recrystallisation or cataclasis. The more uniform grain size of the fine-grained aggregate is usually more typical of the former process. These mantled porphyroclasts may develop wings which allow the sense of shear to be deduced (see section 1:4). Quartz tends to deform in a more homogeneous fashion, tending to wrap around feldspar in a passive manner. At high strains feldspar augen may develop which may be surrounded by fine grained aggregates of feldspar.

At medium to high grades both quartz and feldspar start to behave in a ductile fashion, both deforming by dislocation creep. Both phases tend to have older more deformed grains which grade into subgrains and then into new crystallised grains. At these grades the strength contrast between these two minerals has considerably diminished (see figure)

At low to medium grades (350-500°) at moderate to high strain rates feldspar tends to show "core and mantle" structures whilst quartz tends to show more homogeneous flattening. This is due to slightly varying deformation mechanisms within these two minerals.

Feldspar:- dislocation climb is quite difficult under these conditions so deformation is achieved by GBM recrystallisation accommodated dislocation creep (Passchier & Trouw 1996). Hence the new grains are quite soft due to them being free of

Temperature °C	Very low grade (0, 100, 200, 300)	Low grade (300, 400)	Medium to high grade (400, 500, 600, 700)	Very high grade (700, 800)
Quartz	**Brittle fracturing, pressure solution and solution transfer of material are dominant deformation mechanisms** Fractures in grains, evidence for pressure solution and depostion of material, undulose extinction. Depostion of material sometimes in veins.	**Dislocation glide and creep become important** "Sweeping" undulose extinction, deformation lamallae and subgrain formation	**Dislocation creep is dominant with prism {m}<a> slip becoming important. Recovery and dynamic recrystallisation (SR) become important although with higher temperatures recrystallisation mechanism changes to combined GBM and SR processes.** Old flattened grains with subgrains which grade into smaller recrystallised subgrains. Oblique foliations may develop due to the operation of GBM and SR regimes	**Rapid recovery and recrystn. Prism slip {m}<c> becomes important. Secondary grain growth becomes important** Extensive recrystallisation with grains having a strain free appearance.
Feldspars	**Deformation by brittle fracturing and cataclastic flow** Structures include angular grain fragments which have variable grain size. Such grains show internal deformation, i.e. grain scale faults, bent cleavage planes and undulose extinction (due to sub-microscopic fracturing).	**Internal fracturing assisted by minor dislocation glide** Flame perthite, bent twins and tapering deformation twins. Undulose extinction	**Dislocation glide possible and recrystn. (SR) become important around margins. Microkinking common** "Core and Mantle" structures "bookshelf" fractures **Dislocation glide and climb become easier. Both GBM and SR recrystallisation are occurring. Myrmekite in K-feldspar common** Core and mantle structures still common although boundary between core and recrystallising is less prominent. Myrmekite is common in K-feldspar on faces normal to maximum stress direction. Fracturing uncommon with flame perthite also absent.	
Micas	**Micas deform only by slip on (001)<001> or(001)[100]** Abundant evidence of accommodation mechanisms such as pressure solution and fracturing, kinking and folding	Biotite behaves ductilely above 250°C Muscovite is more resistant to deformation and forms "mica-fish"	**Grain Boundary migration recrytsallisation becomes important at medium to high grades**	

Figure 1:4:- Behaviour of common minerals in response to deformation at varying temperatures. (Compiled from Passchier & Trouw (1996) and Stewart (1997). The text in bold describes the processes taking place; text in plain describes the textures formed by such processes. N.B. An increase in fluid content and lowering of strain rate has the similar effect of raising the temperature.
(The mottled shading represents the onset of ductile defomation processes within the respective minerals)

dislocation tangles and when they do develop tangles GBM can easily replace them. Therefore the mantle tends to be more soft than the core with deformation concentrating within the mantle. With progressive deformation the mantle of the feldspar will grow at the expense of the core.

Quartz:- dislocation creep is mainly accommodated by dislocation climb and SR recrystallisation and hence the new grains still have the same dislocation density as the old grains. The similar strengths of the old grains and new recrystallised grains means the rock will deform in a more homogeneous fashion with no core and mantle structures developing.

During the cooling of a deformed granite one would expect the high temperature deformation fabrics in figure 1:4 to be progressively overprinted by lower temperature fabrics. Figure 1:3a shows an idealised sketch of magmatic fabric overprinted by solid-state fabrics where quartz is typically lenticular (contributing to the formation of a foliation) and displays strong undulose extinction. Feldspar may also show evidence of intracrystalline deformation. This type of fabric is termed a Crystal Plastic Strain (CPS) fabric (Hutton 1988a), equivalent to "solid-state deformation" fabric of Blumenfeld & Bouchez (1988). CPS fabrics form at moderate temperatures (350-450°C) where deformation is essentially homogeneous, whilst at lower temperatures strain becomes heterogeneous with the development of grain scale shear bands (S-C fabrics). Ultimately if heat is still high enough deformation may be confined to zones of high strain (often anastomising around low strain zones) leading to the ultimate development of mylonites (Gapais & Barbarin 1986; Gapais 1989).

1:3:4 A Note on Cloosian Granite structures

Cloos (in Balk 1937) divided structures seen within granites into two main types which formed the basis of structural classification in igneous rocks for many years (e.g. Balk 1937). Here Cloosian ideas on granite structures are compared and contrasted with the recent classification of such structures described in the previous section. The two main types of structure were, as described by Balk (1937) (*in* Berger & Pitcher 1970):-

a) *Primary Structures*

Structures that develop "during the time of consolidation" and includes the free rotation of suspended particles into alignment in a melt. This feature is analogous with PFC fabrics, although Cloos believed this alignment was due to "flow" of the granite against the wall caused by so-called magmatic currents which are comparable to water in a stream. Balk (1937) also claimed that the presence of inclusions aligned parallel to the flow plane was a classic example of primary flow. Joints were also classed as primary structures, especially in the case where their orientations displayed a geometric relationship to the flow alignments. Balk (1937) claimed that such joints allowed "continuation of the viscous elongation" for the cooling magma.

b) Secondary Structures

These were structures that developed below the solidus temperature of the magma which were believed to be essentially metamorphic in origin and included the formation of foliations which commonly cross-cut internal contacts.

The views of Cloos and Balk were challenged by later works of Berger & Pitcher (1970); Pitcher & Berger (1972) and Hutton (1988a) saying granitic melts are not analogous to water, instead being a viscous material capable of being deformed, although the philosophy of primary igneous flow has continued as recent as Marre (1986) (as described in section 1:2). Berger & Pitcher (1970) stated that sedimentary rocks could be folded without signs of macroscopic fracture and should hence be treated as material which could flow in a plastic fashion. The same authors state that inclusions aligned parallel to flow planes are not the result of primary flow but can form in unequivocally metamorphic conditions, e.g. boudins. The presence of cross-cutting foliations is also not necessarily a metamorphic feature but can form during the crystallisation history when ductility contrasts are reduced, a feature Pitcher & Berger (1972) observed in the pluton which forms the basis of this authors research. This is where the primary and secondary structures of Cloos break down because if there are either external tectonic forces or internal buoyancy forces acting on the granite then such discordant foliations may form. It is now generally accepted that so-called "flow" within granites is related to the above mentioned forces and is not the result of magmatic currents. It was for this reason why Hutton (1988a) claimed the terms "primary" and "secondary" should be abandoned and replaced with the less genetic classification described in sections 1:3, i.e. PFC and CPS fabrics. Finally Berger & Pitcher (1970) stated the analysis of joint orientations in ascertaining tectonic models was of limited use as it was difficult to separate primary joints from those formed by pressure release during uplift and erosion.

1:3:5: Strain Analysis

The analysis of strain is a useful method of determining the strain magnitudes and estimating the orientation of the stresses acting up on a granite body and its wall rocks during its emplacement and cooling history. Collection of data from a series of locations within a pluton can lead to an understanding of the deformation which a pluton may have witnessed. The strain ellipsoid of Flinn (1965) is a method of representation which records the relative amounts of deformation in respect to three mutually perpendicular axes of a presumed initial spherical object. The X, Y and Z principal strain axes correspond to the stretching direction, the intermediate stretching direction and flattening direction respectively. In granitic rocks the qualitative approach is to determine the presence of a foliation and combined lineation. This can be related to the strain ellipsoid as S, L and LS tectonites. L-type linear fabrics record mainly constrictional (stretching) prolate strains (where $X \neq Y = Z$); S-type planar fabrics record flattening, oblate strains (where $X = Y \neq Z$) and the generally more common LS-type fabrics which form through plane strain ($X > Y > Z$, with $Y = 1$) where stretching in X is reciprocated by flattening in Z). Pure L-type and pure S-type fabrics are end-members whilst LS types are intermediate between the two (figure 1:5).

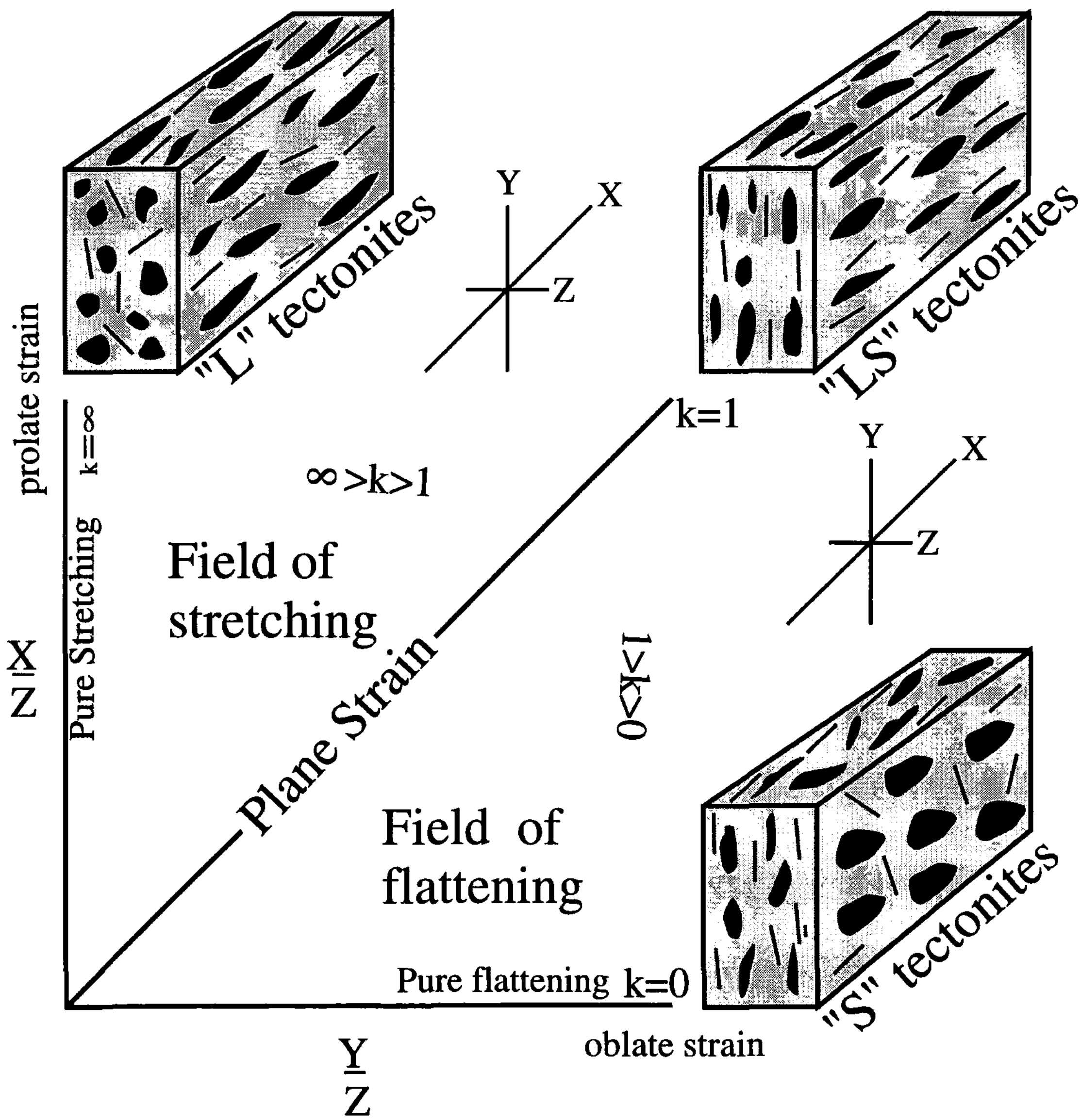


Figure 1:5 Flinn Diagram:- measurements of deformed enclaves within granites can be used to calculate k-values which allow the shape of the strain ellipsoid to be determined, (modified from Hutton 1988).

The measurement of the orientation of these planar and linear features over the whole area of the pluton and its wallrocks gives an insight into the time averaged strain magnitudes and the likely stress regime that acted upon the pluton.

A more quantitative approach is through the measurement of enclaves shape ratios within granitic rocks. "Mafic" enclaves, which are composed of micro-diorite, micro-tonalite and micro-granodiorite, are used as they are believed to have an approximately original spherical shape and tend to give more reliable results than measurements carried out on country rock enclaves (Hutton 1982a, 1982b). Measurement of the length to width ratios on the XY, XZ and YZ planes can lead to the estimation of strain magnitudes and a K-value for the strain ellipsoid which can then be plotted on Flinn plot (after Flinn 1965). For the purposes of collection of data the natural surfaces of erosion often lie parallel or sub-parallel to the principal planes of the strain ellipsoid (Hutton 1988b). This is most often the plane parallel to the foliation, but perpendicular to lineation (XZ) and the plane perpendicular to both the lineation and foliation (YZ) (Ingram 1992).

1:4 Shear Sense Indicators

Non-coaxial deformation is observed to be a very common way in which rocks are deformed. Simple shear and transpressional shear are examples of this type of deformation which tends to produce asymmetric structures in rocks due to the rotation of the finite strain axes with time. This allows the sense of movement, (kinematics) to be deduced. Such structures are termed shear sense indicators. Within tectonites the shear sense indicators should be observed on surfaces parallel to the lineation but perpendicular to the foliation (XZ plane of the strain ellipsoid). When studying granite bodies it is important to study both the shear sense indicators within the wall rock and the granite, as it cools, with the former recording all the strain within the solid-state, whilst the granite may record deformation in the magmatic state and then into the solid-state. Shear sense indicators which develop in the magmatic state will be discussed first with the main solid-state fabrics which are commonly seen in granitoid rocks discussed after.

1:4:1 Magmatic State

Blumenfeld & Bouchez (1988) discussed the criteria which enable the determination of magmatic state shear sense. These include:-

i) **C-S' obliquity**:- the obliquity of early aligned phenocrysts to the shear plane. For example in a granite vein which has sliding walls the phenocryst will be rotated into the XY plane (figure 1:6a). In a vein with non-sliding walls and magma flow along

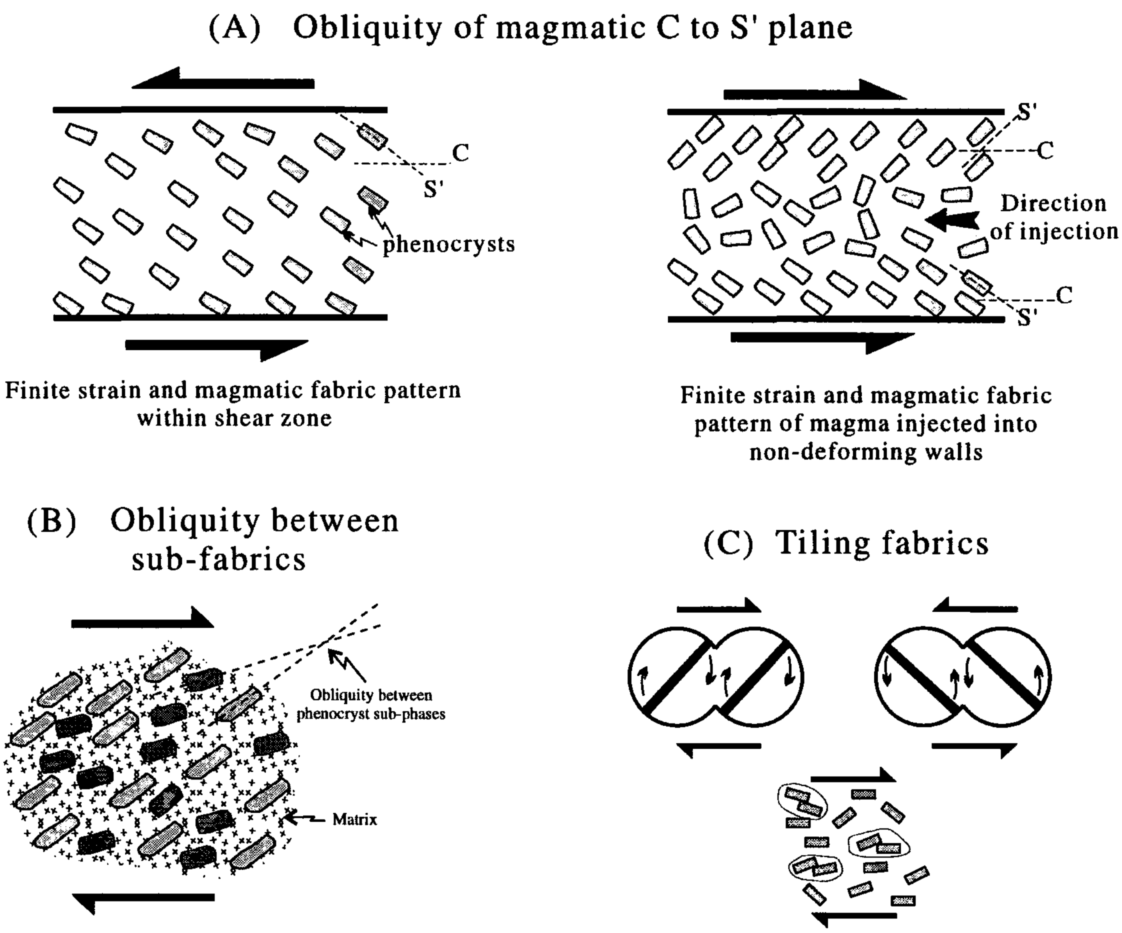


Figure 1.6 Magmatic state structures in granitoid magmas; (a) C-S' (fabric/wallrock) obliquity; (b) Phenocryst sub-phase obliquity; (c) Tiling of phenocrysts. (a, b & c from Blumenfeld & Bouchez 1988).

the vein the sense of shear implied by the phenocryst alignment will be the opposite relative to the medial plane of the vein.

ii) Obliquity between sub-fabrics:- the rotation of mineral phases which have different aspect ratios in a viscous melt will cause obliquity between these phases when subjected to non-coaxial deformation. This is due to the phases with lower aspect ratios rotating more rapidly than the minerals with higher aspect ratios. The difference in rotation rates allows the sense of shear to be determined, (figure 1:6b).

iii) Tiling of megacrysts:- this involves the rotation of relatively closely spaced megacrysts (phenocrysts) within a viscous melt until they touch. This relationship is similar to tiles on a roof (see figure 1:6c). Feldspars which most often show such a relationship. Statistical sampling of populations of tiling needs to be carried out as there are often some numerically subordinate megacrysts tiled in an antithetic sense, (Blumenfeld & Bouchez 1988). For tiling to occur one would expect the crystal percentage to be approaching the RCMP (of Arzi 1978) where rotating crystals start to interfere with one another.

iv) PFC "lock-up" shears:- these have already been described in section 1:3:2. The sense of shear is determined by the sense of rotation of the phenocryst into the shear plane (figure 1:3b) (Hutton & Ingram 1992).

1:4:2: Solid state

Within granitic rocks solid-state shear sense indicators tend to be more common than magmatic features as it is relatively common for granites to deform beyond the RCMP below solidus temperatures. Furthermore if the granite has been subjected to external tectonic forces then it will be common to see similar structures within the wall rock.

1:4:2:1 Common shear sense indicators

Within a simple shear zone the foliation develops at 45° to the shear zone boundary. During progressive deformation this foliation becomes rotated into sub-parallelism with the shear plane. This occurs where the strains are higher, usually in the central part of the shear zone. Therefore across the width of a shear zone one will observe a sigmoidal orientation of the foliation, and this allows the sense of shear to be determined (as in figure 1.7a)

S-C fabrics:- these common indicators within granitoids were first documented by Berthé *et al.* (1979). Related to these in geometry and shear sense are extensional crenulation cleavages and shear bands. These types of structures have been extended into micaceous metamorphic rocks ("mica fish") by Lister & Snoke (1984). The "S" surface refers to "schistosité" and is interpreted as the main foliation within the granite occupying the approximate position of the XY plane in the finite strain ellipsoid (Berthé *et al* 1979). The "C" surfaces ("cisaillement") are small-scale shear planes which deflect the S surfaces producing sigmoidal shapes and this allows the movement along the C planes to be deduced. The "C" plane, in true S-C fabrics, are taken to lie parallel to the shear plane of the progressive deformation with the sense of shear synthetic to the main shear zone (Berthé *et al.* 1979). Both S and C planes can form from the onset of deformation with their formation thought to be broadly synchronous (Platt 1984). The sigmoidal shape of the S planes is produced by the progressive back rotation of the finite strain axes in response to increasing shear strain and also from deflection due to movement along the C planes (Platt 1984; Hamner & Passchier 1991) (figure 1:7b).

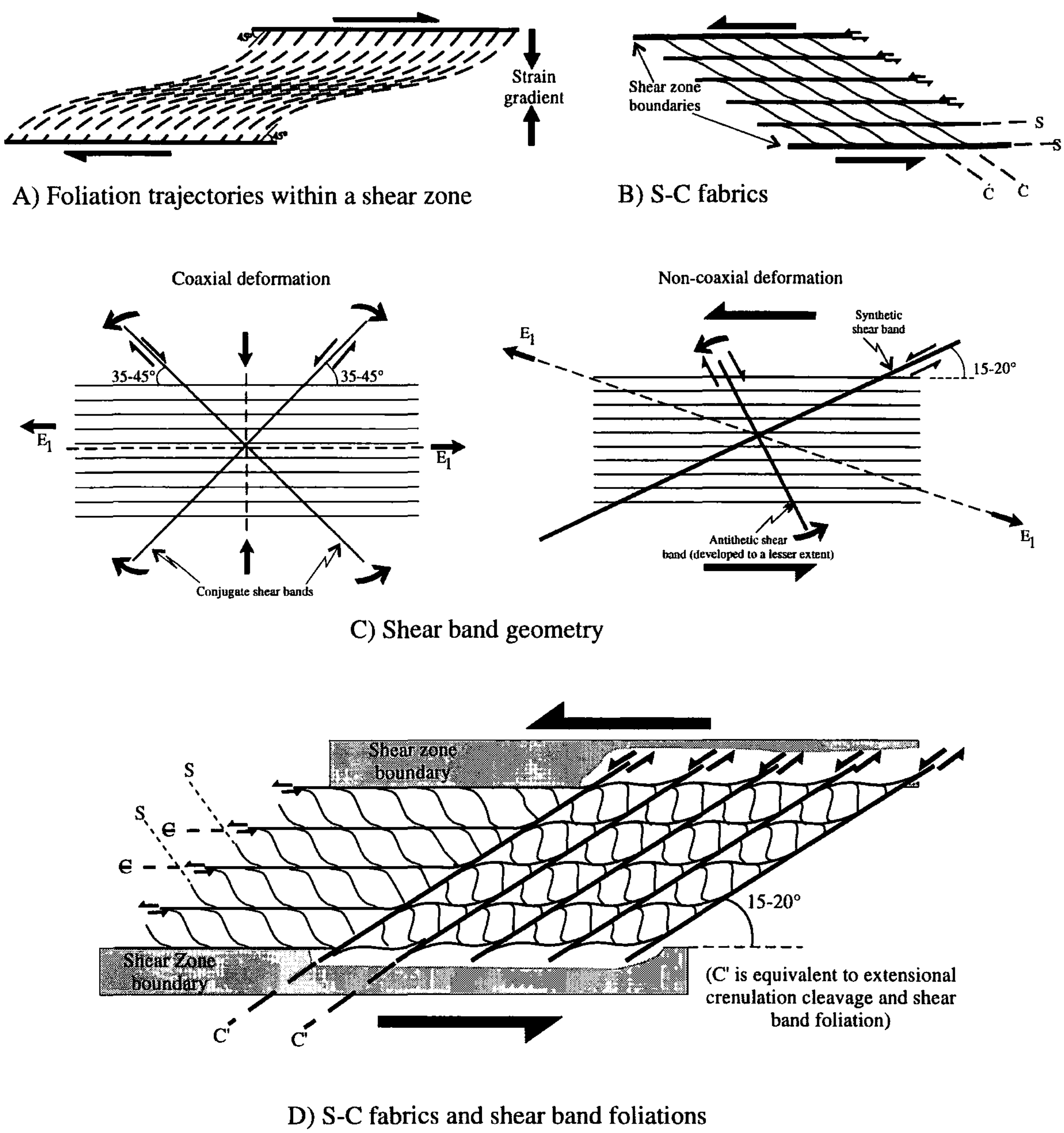


Figure 1:7 (a) Strain variation within shear zones, (after Twiss & Moores 1992) (b) S-C fabrics (Berthé *et al.* 1979). (c) Extensional shear bands (Platt 1984). (d) Relationship of S-C fabrics to shear bands and extensional crenulation cleavage, (C')

The geometry of S-C fabrics is identical to that of extensional crenulation cleavage (also called S-C' fabric; this term is used in this thesis) and shear bands (figure 1:7 c & d) but these are believed to form at a later stage of the ongoing deformation, or they may well form in an unrelated phase of deformation (Platt 1984). Extensional crenulation cleavages are essentially small scale shear zones (like the C fabric) which, unlike the C planes, form at oblique angles to the main boundary of the shear zone. (Platt 1984). This type of cleavage sometimes occurs in conjugate sets and in coaxial regimes where both sets may be equally developed whilst in non-coaxial regimes the antithetic set is developed to a much lesser degree (e.g. Hutton 1977; 1982). In the latter regime the rotation of the antithetic set into orientations unfavourable for slip is the reason for its lower abundance. The more dominant synthetic set initiates at between 15-20° to the shear zone boundary and hence gives the sense of shear. With high shear strains the earlier formed extensional crenulation cleavages rotate towards the shear plane. This can then lead to the development of new sets which overprint the earlier ones (Platt 1984; Hamner & Passchier 1991). Figure 1:7d shows the schematic relationship of C' planes overprinting earlier S-C fabrics

Other shear sense indicators (*for a full review see* Hamner & Passchier 1991) include the effect of "rigid" inclusions (porphyroclasts) in a relatively ductile matrix. Two types of inclusion have been documented by most workers, (e.g. Passchier & Simpson 1986) with the sub-division based on whether or not the inclusions have been rotated. Sense of shear is determined by the common presence of "winged" appendages on the inclusions with the asymmetry of these "wings" giving the sense of shear. The two types are (see figure 1:8 a & b):-

- σ-Type porphyroclasts
- δ-Type porphyroclasts

In granites the porphyroclasts are composed of feldspar, whilst the "wings", which form parallel to the foliation plane, are commonly composed of fine-grained recrystallised material which has the same composition as the host or is composed of reaction softened material, such as quartz and white mica, resulting from the breakdown of feldspar. Pressure shadows and pressure fringes (often composed of quartz or phyllosilicates) which precipitate in low pressure zones adjacent to "stiff" inclusions have a similar geometry to σ-porphyroclasts and can be used in the same way to gain the sense of shear. δ-porphyroblasts have similar wings but the inclusion rotates with the same sense of vorticity as the shear zone.

Porphyroblasts may grow within a deforming matrix and may contain inclusion trails of the external fabric which often get rotated in the porphyroblast as it grows in a manner similar to the formation of a rolling snowball (figure 1:8c). Typical porphyroblasts are garnet and staurolite and are useful for constraining the timing of deformation in the wall-rock with that in the pluton.

These shear sense indicators form when the material is deforming in a ductile manner. Depending on temperature, pressure and strain rate some mineral phases may deform in a brittle fashion (e.g. plagioclase) whilst other co-existing minerals, such as quartz, deform in a ductile manner. Therefore the presence of microfractures within brittle phases can be used for shear sense indication. "Domino" structures (figure 1.8e & f), although cited as a good indicator of shear sense, these structures can be formed by both senses of shear so should only be used when allied with other shear sense structures (Hamner & Passchier 1991).

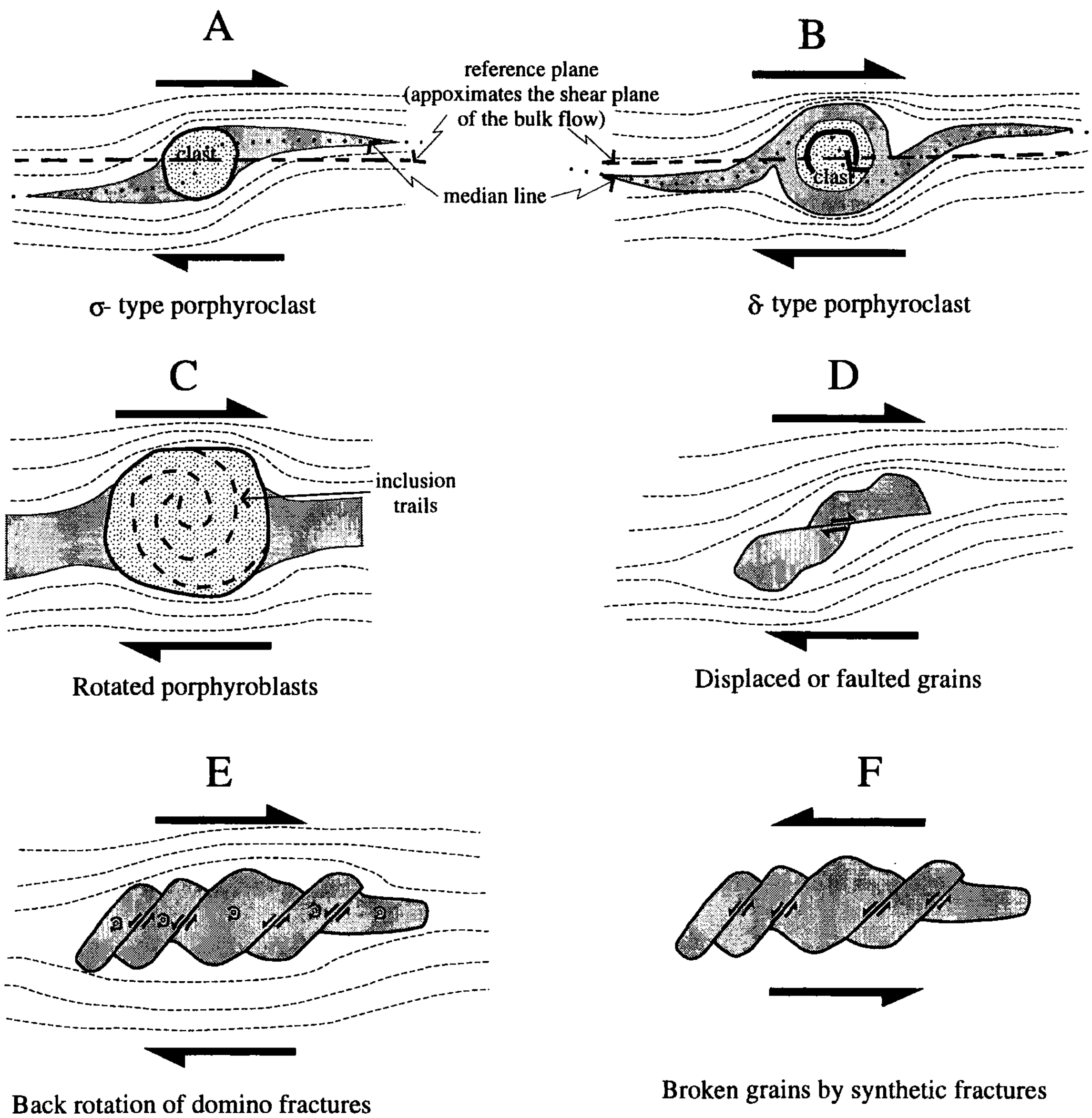


Figure 1:8 Common shear sense indicators, (a, b, c, d & e from Twiss & Moores 1992).

Crystallographic fabrics can be used to determine the deformational regime that a rock has undergone. In quartzo-feldspathic rocks quartz (a-axis and c-axis orientations) is most commonly used to differentiate between coaxial and non-coaxial deformation. (Gapais & Cobbold 1987). The biaxial nature of biotite and plagioclase makes orientation methods more difficult so they tend not to be used as much.

Veins form useful indicators with their development related to the orientation of the instantaneous tensile stress direction ($\sigma3$). As the vein propagates the tips open perpendicular to $\sigma3$, whilst the central portions may get rotated in the same direction as the bulk shear (Ramsay & Graham 1970), ultimately producing sigmoidal shaped veins (figure 1.9b). En echelon arrays of veins are a relatively common feature within brittlely deformed rocks. The orientation of fibres within veins can be used to deduce whether or not it has opened orthogonally or obliquely (figure 1:9c). Another use of veins (quartz veins or pegmatite) is where they have been folded or boudinaged in association with the development of the foliation. This allows the orientation of the finite strain ellipsoid to be established. This method can be adapted to distinguish the difference between non-coaxial and coaxial deformation. During progressive deformation there will be an overlap between the extension and shortening fields of the finite and instantaneous strain ellipsoids (figure 1.9a). For coaxial deformation the overlap will be symmetric, whilst for non-coaxial deformation the overlap is asymmetric allowing the sense of shear to be determined (Hutton 1982).

Deflected marker horizons (e.g. in granitoid rocks features such as banding, dykes and veins) give reliable shear sense and also allow the magnitude of displacement to be determined.

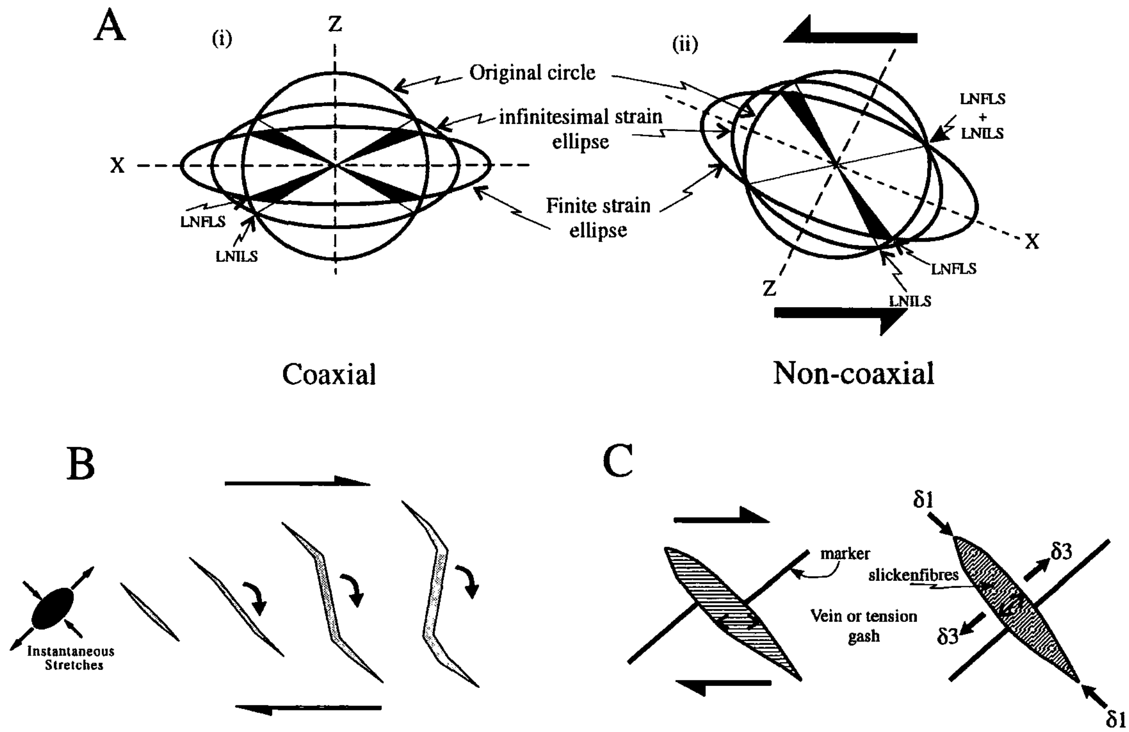


Figure 1:9 The use of veins in determining shear sense; a) using veins and dykes to distinguish between coaxial and non-coaxial deformation (after Hutton 1982); b) vein growth and orientation (Ramsay & Graham 1967); c) obliquely and orthogonally opened veins.

1:5 Mechanisms of Granite Emplacement

The methods described above concerning rheology and the structures which develop within granitic rocks can be applied in the field with the aid of detailed mapping and data collection to ascertain a model for the emplacement of granitic plutons.

In recent years the timing of regional or local deformation with respect to intrusive events has become of particular importance. One of the first applications to this type of work was performed by Pitcher & Read (1959) and Pitcher & Berger (1972) on the Main Donegal Granite, NW Ireland where the deformation in the granite was correlated with the deformation in the aureole and beyond. This work was extended by Hutton (1977) where the deformation within the granite and its adjacent wallrocks was incorporated into a well established regional chronology of deformation throughout NW Donegal. Once such a methodology has been applied then radiometric dating of granite plutons can help to constrain the age of regional deformations. This method has its pitfalls particularly if the phases of deformation have not been correctly identified, e.g. the age of the Ben Vurich Granite of Scotland and its constraints on the age of Dalradian regional deformation is a case in point (*see* Tanner (1996) *and references therein*). Furthermore, if a granite body lies within a tectonic structure, such as a shear zone, and is synchronously deformed during its cooling history (i.e. syn-kinematic) then dating can constrain the age of movements along such tectonic features. Improved methods of dating, which are not likely to be reset by subsequent deformation and metamorphism, have greatly improved this line of study (e.g. U/Pb dating of zircons by SHRIMP (Sensitive high resolution ion-microprobe).

Two main sub-divisions of granite emplacement exist.

1:5:1 Forceful Emplacement

Forceful mechanisms imply the mechanical distortion or displacement of the country-rock to produce space for granitic magma. Throughout the world there are plutons which show the following characteristics of forceful emplacement: circular to sub-circular outcrop shapes, steeply inclined deformation fabrics within the adjacent country rocks and granite, which are generally concordant and intensify towards the granite-countryrock contact (but may not always be present in the latter (England 1990)), distortion or deflection of pre-existing structures in the wall rocks. These features have been observed around many granites (e.g. Arran (England 1990, 1992; Ardara, (Holder 1979) and Canniball Creek (Bateman 1984b). The structures developed above imply the granite has deformed its country-rock and often itself. The following discussion will address how such processes may occur.

Diapirism

Many of the above structures have been attributed to diapiric ascent of granitic magmas. The common occurrence of similar structures during analogue modelling of diapirism has been the main reason for this type of deformation being believed to have been caused by diapirs. Diapirs are forceful magmatic bodies which rise through the crust due to buoyancy and have the ability to plastically deform the surrounding country rocks. During diapiric ascent one would expect to see complex strain patterns developing around the pluton, i.e. ahead of the pluton there would be strong flattening strains; on the flanks there would be simple shear strains with vertical stretching; whilst in the tail there would be constrictional strains with vertical stretching (Cruden 1988). Such structures or complex strain patterns have rarely been observed in natural examples.

Ballooning

Detailed structural studies of many of the so-called classic diapirs, e.g. Ardara, (Holder 1979), Chindamora (Ramsay 1989) and Papoose Flat (Brun et al 1990) have shown that the strain around these plutons show uniform flattening strains. These strains have been attributed to in situ ballooning of a granite where at the level of emplacement the granite expands in a fashion similar to an inflating balloon. In situ expansion may be produced by in situ expansion of multiple magmatic batches where the latter batches tend to deform the earlier ones, or by "auto-intrusion" where the more buoyant lower portion of a diapir intrudes into the cooler upper part when ascent is arrested (England 1990). The way in which one distinguishes between ballooning and diapirism has been a topic of considerable debate (Bateman 1984, Pitcher 1993; Clemens & Mawer 1992). The movement of a diapir through the crust is by upward movement, therefore one might expect to see vertical cylindrical shear zones around the diapir which show pluton up displacements. Around the equatorial regions of a pluton (the identification of which is not always easy in the field) one would expect to see upturned strata. Furthermore below the equatorial levels of a diapir one would expect oblate strains that have been overprinted by more prolate strains. Only a handful of plutons show such structures e.g. Criffel (Courrioux 1987) and North Arran (England 1990, 1992).

1:5:2 Passive emplacement

This type of emplacement is typified by the absence of deformational structures within the granite and its wall-rocks. The two classic mechanisms of passive emplacement are stoping and cauldron subsidence. The common occurrence within plutons of angular fragments of locally derived countryrock implies the intruding granite has exploited fractures within the countryrock. The angular nature implies the country rock is brittle, a feature typical of the upper crust. The stoped blocks then subside into the magma allowing magma to rise vertically. The process then repeats allowing progressive ascent of the granite magma. Despite the common occurrence of stoped blocks within plutons e.g. Thorr and Fanad plutons, Donegal, (Pitcher & Berger 1972) stoping is not regarded as an important mechanism in ascent and emplacement. Firstly there is a space problem where the stoped blocks will rotate as they fall through the magma and hence will take up more space. Also to remove the stoped blocks the pluton must get wider at depth otherwise "block-choking" will occur preventing the magma from rising. Such plutons which get wider at depth are not seen. Furthermore the presence of stoped blocks increases the surface area of the countryrock resulting in greater thermal conduction and hence freezing of the magma. Therefore stoping tends to be a more localised phenomena occurring for example in the roof zones of plutons (Hutton 1997 *in press*).

Cauldron subsidence and the formation of granite ring complexes is another method of passive emplacement where the Earth's surface behaves as a free surface, i.e. stresses will be resolved parallel and perpendicular to the Earth's surface and also this surface may be displaced vertically to allow space to be created below (Hutton 1997 *in press*). The above principles have been applied by Anderson (1936) and Roberts (1970) to granite bodies at high levels in the crust. The presence of excess magmatic pressure in a pluton at high levels in the crust may cause uplift of the free surface by concentric inwardly dipping fractures which may then fill with melt from below, hence producing cone sheets. The drop in pressure due to release of magma will cause the overlying rocks to collapse and fracture producing outwardly dipping fractures which will subsequently fill with magma to produce ring dykes. Commonly ring dykes have flat tops and steeply inclined sides. The repeated intrusion of compositionally dissimilar granites in this fashion leads to the development of centred complexes when viewed in outcrop e.g. Barnesmore (Walker & Leedal 1954); Slieve Gullion (Richey 1932) and the ring complexes of northern Nigeria (Pitcher 1993).

1:5:3 Tectonic Controls on granite emplacement

The relationship between passive and forceful emplacement is often obscure (Hutton 1988). It is accepted that granites rise through the crust due to buoyancy (the density differential between the magma and the rock column through which it is rising) which implies a forceful nature. Therefore at lower levels in the crust, below the level of neutral buoyancy (LNB) one might expect forcefully intruded granites whilst at higher levels in the crust, at and above the LNB, where the surface of the

Earth behaves as a free surface one would expect emplacement by passive mechanisms such as cauldron subsidence to occur. The above information relates the depth within the crust to the level of emplacement, a feature which Read (1957) & Buddington (1959) drew attention to. These authors stated the presence of bimodally emplaced plutons at the same level within the crust was due to time within a batholith, where the forcefully emplaced granites of the lower crust had been uplifted and eroded and later passively intruded by granites in the now upper crust: in essence the "Granite Series". As already mentioned the presence in both space and time of bimodally intruded granites in the Donegal Batholith, implied that depth was not the controlling factor on emplacement mechanism and hence the "Granite Series" model was not applicable.

The tectonic approach of Hutton (1988*a*) stated that the variety of emplacement mechanisms observed in granite plutons is a "..*function of the interaction between natural magma buoyancy forces and ambient tectonic forces*", i.e. the style of intrusion is due to the interaction of buoyancy and tectonic forces. Passive plutons were emplaced when the extension rate of the cavity exceeds the internal compressive buoyancy forces. In contrast; forceful plutons will be found when the extension rates in cavity opening are less than the compressive magmatic buoyancy forces. Figure 1.10 shows that the above mechanisms are end-members and one would expect to see plutons showing a combination of forceful and passive mechanisms, (Hutton 1988*a*; Jacques & Reavy 1994).

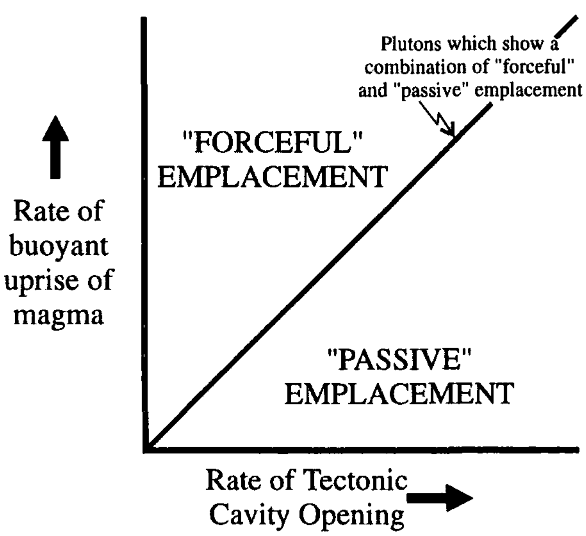


Figure 1:10
The relationship between external tectonic forces and internal buoyancy forces on governing emplacement styles (*after* Hutton 1988).

Therefore the structural study of a granite and its wallrocks should attempt to evaluate the role of tectonic controls and buoyancy forces during emplacement. Hutton (1982) demonstrated that the intrusion of the Main Donegal Granite was synchronous with deformation in features such as active faults or shear zones, i.e. a syn-kinematic pluton. The list below describes some of the important criteria which enable the identification of such plutons:- i) elongate shapes, where the long axis coincides with belts of deformation and metamorphism (shear zones). ii) drop in strain magnitude when crossing the countryrock to pluton contact as indicated by strain markers in granite and wallrocks. iii) PFC and CPS fabrics, plus lineations which are coaxial and coplanar to equivalent fabrics in the wall-rocks. iv) the growth of syn-kinematic metamorphic minerals in wallrocks (identified by fabric analysis), and v) truncation of shear zone fabric by intrusives which later become deformed by the same progressive deformation (Hutton 1982; 1988*a*; Ingram 1992). The interference of buoyancy related forcefully emplaced plutons and external tectonic forces may lead to the development of foliation triple points. This feature may be located inside or outside the pluton depending on the intensity of the respective strains, (McCaffrey 1989).

The above structures have been identified in plutons emplaced in shear zones in three tectonic regimes:- transcurrent shear (includes transtension + transpression), contractional shear (thrust sense) and extensional shear. Within all three of these tectonic settings there is a need for space creation, e.g. in fault jogs, restraining bends and releasing bends, rampflats and ramps (Sylvester 1988: Sanderson & Marchini 1984). Work by Jacques & Reavy (1994) have shown that, within individual plutons in a shear zone, the orientation of the instantaneous stress axes will govern the distribution of zones of extension and compression throughout the pluton. This, in turn will affect the style of emplacement, i.e. some parts of the pluton display passive and forceful intrusion respectively.

Within the three main tectonic settings many of the plutons mentioned below have been emplaced in the form of sheets of varying thickness and overall shape. Hence large plutons appear to be constructed by the incremental emplacement of smaller sheets or pulses. Aspects of sheeting will be returned to later in this chapter.

a) Transcurrent settings:- as Hutton (1992) points out models for the emplacement of plutons often bear a relationship to smaller scale structural geometries, e.g. vein arrays. Emplacement in tension gash systems (Extamadura (Castro 1986); pull-aparts (Mortagne, (Guinberteau et al 1987); shear zone terminations (Strontian (Hutton 1988a); Solli Hills, Nigeria (Ferré *et al* 1995); Fanad (White & Hutton 1985); dilation produced by longitudinal shear strain gradients, e.g. Main Donegal Granite (Hutton 1982). Other examples of plutons in transcurrent systems include Iberia (Reavy 1989); Ox Mountains (McCaffrey 1992); Brazil (Tommasi et al 1994). Hutton & Reavy (1992) and D'Lemos *et al* (1992) showed that during transpressional orogenesis crustal thickening may often be considerable and may generate crustal melts at the base of the continental lithosphere which may then ascend the shear zones with emplacement governed by the geometry of the shear zone at higher crustal levels.

b) Extension:- syn-tectonic intrusion has been observed in extensional shear zones where space can be created along features such as ramps. The rapakivi granites of Greenland (Hutton *et al.* (1990); Brown *et al.* (1990) and the Xanthi pluton, Greece (Koukouvelas & Pe-Piper 1991) are such examples.

c) Compression:- the examples in (a) and (b) have shown that the space creation is important in siting a pluton, but space creation is not an essential factor. Of the plutons emplaced in thrust regimes, the Great Tonalite Sill, Alaska is the most spectacular example (Hutton & Ingram 1992; Ingram & Hutton (1994). PFC fabrics show emplacement and shear zone thrusting were synchronous. Therefore buoyancy and "head" forces were capable of overcoming compressive tectonic stresses. Another example of intrusion during contemporary shortening is the Veiga Massif, NW Spain (Roman-Berdiel *et al.* 1995) and the Western Unit of the Central Vosges Massif (Blumenfeld & Bouchez 1988).

1:5:4 Sheeted granite plutons

Within all the three tectonic settings described above there are in examples of sheeted granite plutons. Within the Cretaceous-Tertiary aged Great Tonalite Sill the 800 × 20 km pluton is composed of smaller lensoid bodies, separated by country rock screens. Within some of the individual bodies internal sheeting could be observed by the presence of chilled margins (Ingram 1992, *in* Hutton 1992). In Greenland, Hutton *et al.* (1990) observed, within the normal faults, the presence of sheets up to 500 metres thick composed of early Proterozoic rapakivi granite. Within the Caledonian-aged Ox Mountains Granite of Ireland, which lies adjacent to the assumed trace of the Highland Boundary fault, McCaffrey (1989; 1992) reports sheeting within this 27 × 6 km pluton. Within this highly deformed pluton, the steeply inclined sheets intruded into each other and have approximate dimensions of 300 × 2000 m although there are sheets of smaller sizes.

The above examples display evidence of sheeting based on petrographic variations, but in some plutons which are petrographically more homogeneous the evidence of sheeting is much more subtle with other factors such as "ghost stratigraphy" suggesting sheeted geometries. The Main Donegal Granite pluton is supposedly an example of a pluton which shows so-called "cryptic" sheeting (Pitcher & Read (1957): Pitcher (1970); Pitcher & Berger (1972); Hutton (1992)). Within this elongate body there are a series of strike-parallel inclusions of countryrock and older pluton inclusions (raft-trains) which are sometimes traceable for distances up to 20 km, which preserve the stratigraphy seen in the aureole, (Pitcher & Read 1959; Pitcher & Berger 1972). The latter authors believed the inclusions were dismembered roof septa which separated different sheets of granite. Therefore Hutton (1992) stated that the spacing of the raft-trains indicates the size of first order sheets (0.2-0.5 km). Smaller sheets may be present although Hutton (1992) states that the identification of such sheets within petrographically similar granites may be almost impossible as high ambient temperatures may not allow chilled margins to develop. Furthermore rapid intrusion of sheets with high remaining melt fractions may undergo, after emplacement, mixing or mingling or even partial coalescence to form a homogeneous looking granite (Hutton 1992). In summary then, the presence of an aligned "ghost-stratigraphy" maybe the only clue of a sheeted geometry.

Before going on to describe the aims of this thesis which are concerned with the sheeted nature of the above mentioned pluton, the overall geometry of sheeted plutons will be examined and the matter of how this relates to the ascent of granite magmas through the crust will be discussed.

Sheeted plutons can be thought of as granitic dyke complexes and so it is appropriate to discuss the mechanics of dyke emplacement and how this may be related to emplacement in shear zones.

1:5:4:1 Andersonian dykes

Much of this discussion is taken from Hutton (1992) *and references therein.* Anderson believed that dykes were emplaced by a wedging mechanism where there is slight excess pressure of the magma over the compressive stress of the rock. This causes extreme tensile stresses to develop at the tip of the dyke which can easily overcome the lithostatic strength of a rock resulting in fracture and allowing the dyke to propagate. For propagation to continue there must be either excess pressure at source and also, depending on the properties of the magma, buoyancy (Clemens & Mawer 1992; Petford 1996). The excess pressure is optimised in a regional stress field if the dykes align themselves parallel to maximum compressive stress ($\sigma1$) and normal to least compressive stress ($\sigma3$), (figure 1:11a & b).

In the examples of sheeted plutons in the different tectonic regimes noted above it is apparent that none of the sheets lie parallel to $\sigma1$ (figure 1:11c, d & e) and in the extreme case of the Great Tonalite Sill, the sheets are almost at right angles to σ 1. In the case of the Greenland Rapakivi granites and the Ox Mountains Granodiorite the sheets are oblique to both $\sigma1$ and $\sigma3$. Therefore it seems that the model of Anderson does not always apply. Hutton (1992) suggested that within these plutons the geometries and shapes are unrelated to crustal stress systems and instead are controlled by the geometries of the shear zones they intrude into.

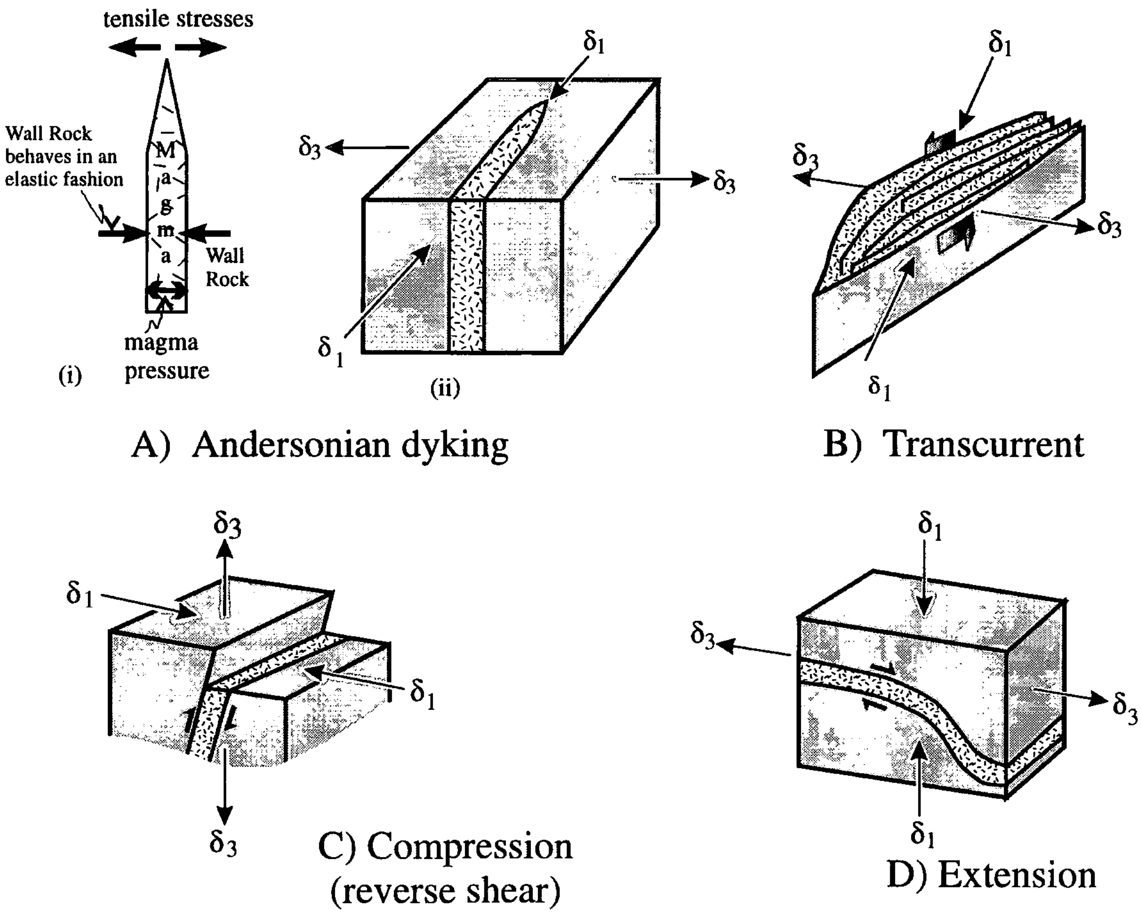


Figure 1:11 The significance of granite sheet orientation to the principal stress axes:- (a & b) Andersonian dyking; (c,d & e) Non- Andersonian dyking of sheeted plutons in the three main tectonic regimes (Hutton 1992).

Andersonian dyking implies propagation through an isotropic, flawless crust, although in the realities of the anisotropic crust one would assume that magmas would exploit zones of weakness and potential areas of dilation, for example shear zones. Along the median planes of shear zones, where rocks tend to be rheologically weakened by high continuous strains, propagation of dykes would be highly effective (Hutton 1992).

Whilst in general a distinction between should be drawn between ascent mechanisms with emplacement phenomena, Hutton (1992) states that there is the possibility of preservation, within sheeted granite complexes, of ascent mechanisms at the level of emplacement. From the examples cited the presence of granite in shear zones may imply that granite has ascended along such structures. It is commonly known that shear zones can extend to deep levels within the crust and may intercept the source region. Where the pluton decides to emplace (i.e. build) during its ascent may be controlled by tectonic space creation or where the effects of buoyancy and the excess pressure at source diminish. As in the case of the Great Tonalite Sill dilational tectonic space creation is not an essential factor in emplacement.

1:6 A Brief discussion on ascent mechanisms

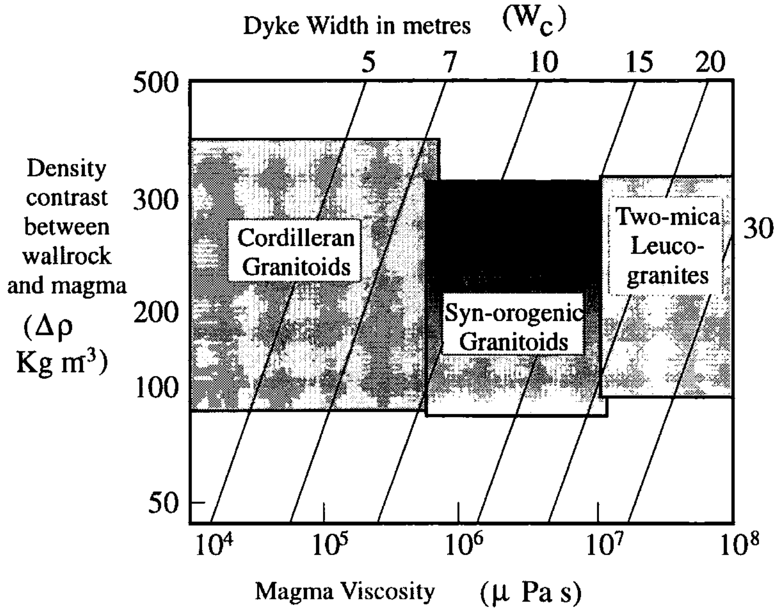
Diapiric rise was recognised as the main ascent mechanism of granitic melts for almost half a century. Diapirs rise due to their internal buoyancy and ability to reduce the viscosity (mainly thermal softening) and push aside country-rocks. This relies on continual convection in the diapir to carry heat to the periphery to heat up the wall-rocks. At lower crustal levels this mechanism is viable where ambient temperatures are high and the countryrocks are rheologically weaker. The development of structures which resembled natural examples in analogue modelling, notably Grout (1945) and Ramberg (1971), and the lack of alternative mechanisms for ascent were the main reasons for diapiric ascent being accepted for so long. It was the during the 1980's that the validity of diapiric ascent at mid to high levels in the crust was first questioned mainly from the work of Marsh (1982) who developed the "Hot Stokes" formula. This states that ascent will cease when the pluton has no longer the heat to reduce the viscosity of its wallrocks (England 1990). Therefore at high levels the cold lithosphere will conduct the heat away too fast resulting in the pluton freezing. Repeated ascent along the same path, raising the ambient temperature, was considered to be essential in allowing diapiric uprise to higher crustal levels (Marsh 1982) which would result in "nested" diapirs at the level of emplacement (Paterson & Vernon 1995). It was at this time when many of the classic examples of "diapirs" e.g. Ardara (Holder 1979), Cannibal Creek (Bateman 1984) and Chindamora (Ramsay 1989) were shown to develop their geometries from in-situ expansion, or ballooning during emplacement and hence were not "frozen" ascent structures. Furthermore many of these plutons were at levels too high in the crust for diapirs to rise to according to the Hot Stokes formula of Marsh (1982).

A current major alternative to diapiric ascent is dyke ascent as first seriously suggested for granite magmas by Clemens & Mawer (1992). In the past it had been thought that granite magmas (even when crystal free) were too viscous to ascend in such narrow structures, being generally 10^3-10^6 times more viscous than basalt.

The width of dykes is related to the viscosity, i.e. the higher the viscosity the wider the dyke has to be. Figure 1:12 shows the width of dykes for different granitoids when density contrast is plotted against viscosity. The critical width values of granitoid dykes, below which freezing will occur, is between $\approx$ 6-30 metres for dykes of 1 km length (Petford 1996). Crystal suspension is an important factor although phenocrysts contents of <35% have effective viscosities similar to crystal free melts. Melts with >50% are to viscous too transport in dykes (Lister & Kerr 1991). In comparison to diapiric rise the flow rates in dykes are magnitudes faster as much as $\approx 10^{-2}$ m/s in comparison to 10^{-8} m/s for diapirs (Mahon *et al* 1996; *in* Clemens & Mawer 1992). Therefore the initiation of felsic dykes could be described as catastrophic. Ascending magma will rise adiabatically from the source and expansion (caused by increase in the partial molar volume of "water") due to decompression further increasing hydrostatic pressure which will cause the dyke to widen and hence create higher tensile stresses at the tip of the propagating dyke. Adiabatic ascent will also cause the resorption of crystal phases plus restitic material within the melt. (Petford *et al.* 1993;1994). A vital requirement for successful felsic magma transport is that the rate of ascent is rapid enough to prevent thermal conduction of heat into the wall rock which would result in solidification of magma and rapid increase in viscosity, (Petford 1996; Clemens & Mawer 1992). Petford (1996) maintains that as long as extraction from source is rapid enough then the majority of crustal contamination will be confined to the site of emplacement.

Clemens & Mawer (1992) stated that felsic melts collect and form dykes by fracture propagation, whilst Petford *et al.* (1993) state that felsic dykes ascend in fault related conduits (Petford 1996) with initiation by crustal extension (Petford 1996). In both methods of dyke ascent the physical behaviour is similar. During partial melting of fluid absent crust there is a large increase in volume, between 2.4-18 vol.% (Clemens & Mawer 1992) produced by the breakdown of hydrous phases such as hornblende and biotite. The maintenance of a solid framework in the source region will increase the pore pressure and hence lowering the effective normal stresses resulting in brittle failure of the source (Clemens & Mawer 1992). The exact method in which melts collect and initiate as dykes is still not fully understood but active research is attempting to shed light on this problem (Petford *pers comm.* 1997). As Rubin (1995) stated the main problem is during the early formation of the dykes at the source i.e. getting the felsic dyke wide enough to prevent freezing. For dyke propagation two forces are required: i) buoyancy. ii) Excess pressure at source (i.e. the pressure in excess of lithostatic load (Petford 1996)). The latter force can cause dyke propagation in any direction whilst buoyancy will cause vertical fractures to develop. During vertical transport the buoyancy force must be larger than the excess pressure, with the latter force being responsible for initiating fracture. After fracturing has commenced the shape and orientation of the dyke will be controlled to a considerable extent by the regional stress regime although final emplacement geometry may differ from that during ascent (Petford 1996 *and references therein*). The most efficient geometry of a dyke is a planar feature.

Figure 1:12
The critical dyke width for granitoid magmas when magma viscosity is plotted against the wallrock/ magma density contrast, (taken from Petford 1996)



The construction of granite batholiths by this method is presumed to be a rapid affair, where felsic melts can ascend through 10-20 km of crust in a matter of months, rather than thousands or millions of years, as expected for diapiric ascent (Petford 1996; Clemens & Mawer 1992). Petford (1996) claims large plutons can be constructed in as little as 10^4 years although this rate depends on the amount and rate of melt extraction at source. The implication is that dyke ascent is not continuous but implies that it is episodic where melt has to accumulate until it reaches a critical value where dyke initiation may occur. Once the source is drained the process repeats the cycle.

Ascent by this method is very attractive especially when one actually observes sheeted granites at the site of emplacement. It may be to early to discredit diapirs as work by Weinberg & Podladchikov (1995); Weinberg (1996) have shown that thermal constraints alone are not the only rate limiting process as believed in the "Hot Stokes" diapirs. Weinberg claims that the mechanism for softening (reducing viscosity) the crust is not thermal, but is strain dependent i.e. highest strains around the pluton will lower country-rock viscosites. The model relies on the diapir velocity to strain soften the country rocks. Their modelling assumes a power law crust and not a Newtonian crust as other models have implied (Weinberg 1996). So-called "power law" diapirs are believed to rise (although they are modelled to sink as well if their density becomes greater than the surrounding wall rock) to depths of between 10-15 km below the surface (Weinberg 1996). Only in rare circumstances such as anomalous heat flows or very low viscosities can power law diapirs rise higher than 10 km below the Earth's surface. An interesting proposal which Weinberg & Podladchikov (1996) and Weinberg (1996) stated is that at high levels in the crust where it is generally "stiffer" and colder the diapir may halt its ascent and give rise to dyke propagation where such dykes will drain the felsic melt from the diapir allowing granite melts to ascend to higher levels.

1:7 The Objectives of this thesis

This chapter up to this point has described the modern ideas on the rheology of granite magmas, the nature and style of its emplacement and an insight to the ascent mechanisms. The aims of this thesis are to study a cryptically sheeted pluton which was emplaced into a transcurrent shear zone using the Main Donegal Granite, NW Ireland, a pluton which is believed to be classic example, as a case study. A basic map and considerable geological data already exist for this well exposed and well known pluton. A basic subdivision of the petrographic facies has been published (Pitcher & Read 1959) and detailed maps are known for only a few very small areas. With the aid of detailed mapping, microscopic analysis and systematic geochemical sampling, the aim is to try and identify in detail the different granite facies and establish their geometries on both a small and pluton scale. This is combined with field observations on cross-cutting relationships attempts to establish a chronology of emplacement of these facies and how they relate in space and time to the building of the pluton and the shear zone of which it is associated. Furthermore this mapping will help to understand how the varying appearance of granite sheeting may give insight into the rheological and environmental controls affecting a pluton during its construction.

Chapter 2

The pre-granite Geology of Northwest Donegal

2:1 Introduction

The rocks of Donegal lie approximately half-way along the Caledonian orogenic belt, traceable from Svarlbad in the north, through NE Greenland and Scandinavia, into the British Isles, extending on into Newfoundland and the eastern side of the North American continent, (Harris *et al.* 1979). Within the British Isles area of the belt there is a change in strike of the orogen, from NNE-SSW in the northern part of the belt, to NE-SW in the American part of the Caledonides. This feature is related to the Labrador-Greenland Promontory, a pre-orogenic feature of the Laurentian margin. (Hutton & Alsop 1996). The rocks of the British and Irish Caledonides are composed of a series of autochthonous and para-autochthonous stratigraphic sequences (e.g. the Moine and Dalradian) and allochthonous, possibly far travelled terranes, (Hutton 1987).

The Caledonian orogeny records the collision of Laurentia, Baltica, Eastern Avalonia and intervening terranes sometime between the mid-Ordovician and the early-Devonian (Soper *et al.* 1992). Donegal forms part of the "Grampian Terrane", bounded to the north by the Great Glen Fault and to the south by the Fair Head-Clew Bay Line, (possibly the Irish equivalent of the Highland Boundary Fault), (Hutton, 1987). The terrane consists of Neo-Proterozoic rocks belonging to the Dalradian Supergroup. The terrane witnessed across-strike deformation between 590-470 Ma, during the Grampian Orogeny, (Robertson 1994; Tanner & Leslie 1994 and Tanner 1996). The orogeny records the collision of outboard arc terranes as palaeomagnetic evidence indicates Iapetus was still wide open at 470 Ma, (Soper *et al.* 1992). In NW Donegal this was recorded in a series of NW facing recumbent folds, thrust over Proterozoic basement, which is exposed to the north of the mainland on Inishtrahull and on Islay: the Rhinns Complex, (Fitches *et al.* 1994). In central and southern Donegal and mid-Ulster there is a series of nappes facing and over-thrusting to the SE, (Alsop & Hutton 1993; Hutton & Alsop 1995). Between 460-410 Ma there was uplift of the Grampian Terrane, but no associated deformation comparable to the thrusting events occurring at this time in the NW Highlands of Scotland, i.e. the Moine Thrust System, ~430 Ma, (Soper *et al.* 1992).

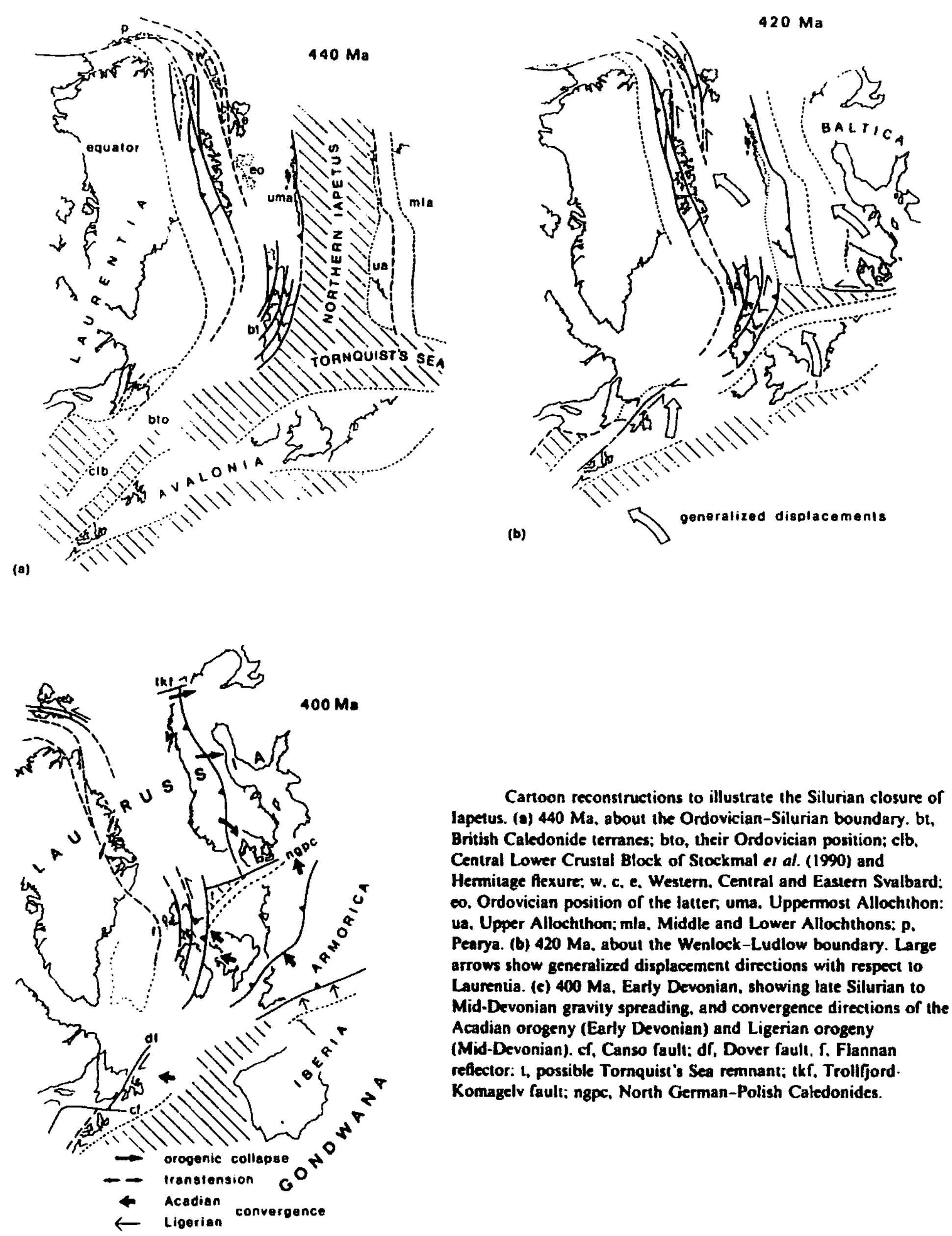


Figure 2:1:- Reconstructions illustrating the closure of Iapetus during the late-Ordovician and Silurian, (after Soper *et al.* 1992)

During the later stages of the Caledonian episode, at 425 Ma and 400 Ma, (mid-Silurian to early-Devonian), major sinistral displacements on NE-SW strike-slip faults and shear zones are recorded, (Hutton & McErlean 1991). Soper *et al.* (1992)

attributed this to the oblique northward impingement of Eastern Avalonia, at 425 Ma and Amorica-Iberia, a fragment of Gondwana at 400 Ma resulting in sinistral transpression. Associated in time with the sinistral displacements are the emplacement of numerous granitic complexes of which the Donegal Batholith is an example, (Watson 1984; Hutton *et al.* 1984; Jaques & Reavy 1994).

The remainder of this chapter will address the geological history of Donegal prior to the emplacement of the Caledonian Granites.

2:2 The Dalradian Supergroup

The Granites comprising the Donegal Batholith (approx. 400 Ma, O'Connor *et al.* 1982) have been intruded into a stratigraphically diverse sequence of rocks belonging to the Dalradian Supergroup, (see figure 2:2). The sediments were deposited on the south-eastern margin of the Laurentian continent, commencing in Neoproterozoic times, (the Appin and Argyll Groups), and possibly continuing into Lower Palaeozoic times (the Southern Upland Groups), Harris and Pitcher (1975). The Appin Group is interpreted as being deposited on a shelf, typified by beach sands, lagoonal facies and darker euxinic mudrocks. This sequence of rocks is interpreted as being deposited whilst active rifting was diminishing, with thermal subsidence becoming more dominant, (Soper & England 1995). The Argyll Group records the deposition of sediments on a more unstable subsiding shelf with polycyclic quartzites and turbiditic type rocks. Major lateral facies changes within this group suggests the development of sub-basins, (Alsop & Hutton 1990). Towards the top of the Argyll Group limestones, turbidites and submarine volcanics are present supporting increasing instability within the Dalradian basin. At the base of the Southern Highlands Group turbiditic facies dominate, containing occasional bands of impure limestone and volcanogenic material. Accompanying this style of deposition was the protracted intrusion of large volumes of tholeiitic magma, notably in Donegal and Argyll, (Tayvallich 595 ±5 Myr, Halliday *et al.* 1989), (Pitcher & Berger (1972); Harris & Pitcher (1975). This volcanism at the base of the Southern Uplands Group is believed to be related to major rifting associated with the opening of the Iapetus ocean with the prograde deposition of turbidites onto newly formed oceanic crust, (Soper & England 1995).

2:3 The Dalradian rocks of Donegal:

The Dalradian of Donegal has been traditionally divided into three major successions, (Pitcher & Berger 1972) which broadly correlate with the major groups of the Scottish Dalradian, (Harris & Pitcher 1975) (see figure 2:2).

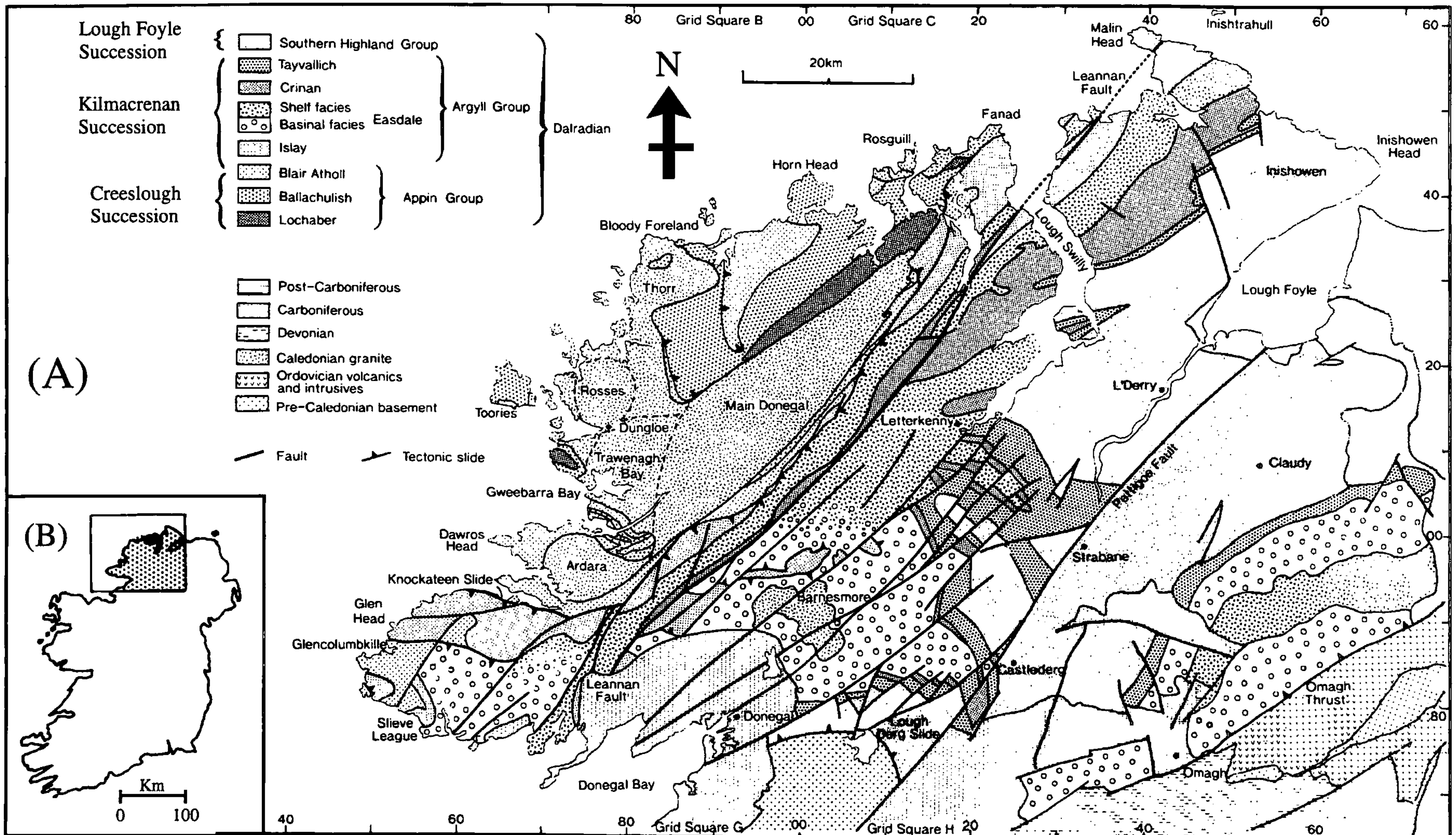


Figure 2:2:- (A) Solid geology of NW Ireland showing the Dalradian at subgroup level (*after* Hutton & Alsop (1996)
(B) Inset of Ireland showing the location of Donegal.

The oldest sequence of rocks belong to the *Creeslough Succession* (Appin Group), into which the most of the Donegal Batholith, (except the Barnesmore Pluton) was emplaced. To the south-east the Creeslough Succession is tectonically overlain by the rocks of the *Kilmacrenan Succession* (Argyll Group), separated by the pre-granite Knockateen Slide. The rocks of this succession show similarities in facies to the Creeslough Succession, although within central Donegal there are major lateral facies changes, implying more basinal facies towards the south. Movement along the Leannan Fault juxtaposes the basinal facies of central Donegal against the more typical shelf facies to the NW of the fault. The clarification of sinistral displacements along the (Devonian) Leannan Fault, (Alsop 1992a) means the basinal rocks of the Slieve League Peninsula were originally just west of Letterkenny, (Pitcher & Berger 1972). (see figure 2:2).

On top of the Kilmacrenan Succession is the turbiditic sequence of rocks belonging to the *Lough Foyle Succession*, (Southern Highlands Group). Again within this succession there are major lateral facies relating to basin architecture during deposition, produced by a major NNE-SSW trending lineament, (Hutton & Alsop 1996). This not only influenced Dalradian sedimentation but possibly the distribution and orientation of Caledonian folds and the siting of the Donegal Granites (see later sections).

2:3:1: The Appin Group of NW Donegal

The Appin Group consists of the Creeslough Succession of Pitcher & Berger 1972) and mainly outcrops to the north-west of the Main Donegal Granite (McCall 1954 and Rickard 1962), extending into Rosguill, (Knill & Knill 1961) and eventually terminating in the Fanad peninsula against the Fanad Pluton and the overlying Kilmacrenan Succession, (White & Hutton 1985). This succession also occurs as isolated exposures separated by granite at Maas; (the "Maas Succession" of Gindy (1953), Iyengar *et al.* (1954), Akaad (1956) and Mithal (1952); along the south-eastern margin of the Donegal Granite complex and to the NW of the Knockateen Slide there is a narrow strip of rocks belonging to the Creeslough Succession, (the so-called "Fintown Succession" of Akaad (1956), Tozer (1955), Cheeseman (1956) and Pande (1954), in Pitcher & Read (1959), (1960), Pulvertaft (1961) and Cambray (1964). These three main areas of outcrop, (i.e. Creeslough, Maas and Fintown) were correlated by Pitcher & Shackleton (1966) creating a unified Creeslough Succession. This was achieved by comparison of lithologies at key localities and tracing the "ghost-stratigraphy" preserved within the Thorr and Main Donegal Plutons which separate the respective areas. Pitcher & Shackleton (1966) divided the Creeslough Succession into nine individual rock formations which had a combined maximum thickness of 4-5 km's. Tectonic thickening and thinning by folding and ductile tectonic sliding respectively results in this value varying from area to area. A full description of the formations which make up the Creeslough Succession is available in Pitcher & Berger (1972), from which the following summary has been taken. Due to the development of an extensive "ghost-stratigraphy" within the Main Donegal Granite the author feels it is necessary to give a brief description of the individual formations, starting with the oldest:- (also see figure 2:3)

The Creeslough Formation:- this outcrops in the low-lying ground immediately to the NW of the Main Donegal Granite, and extends into Rosguill and Fanad. McCall (1954) and Rickard (1962) believed the Lackagh Quartzite, which is in contact with the Main Donegal Granite, to be the oldest member of the succession, though work by Pitcher & Shackleton (1966) on Rosguill showed that this quartzite is a folded repetition of the younger Ards Quartzite, with the Creeslough Fm. occupying the core of an anticlinal structure (the Aghla Anticline of Hutton 1977). The original thickness estimates of McCall and Rickard for the Creeslough Fm. were therefore too large. The formation is composed of non-graphitic, grey-green pelites and semi-pelites intercalated with limestones and calc-pelites. McCall, (1954) identified three prominent limestone bands forming invaluable marker horizons in the moderately exposed ground, (from oldest to youngest: Duntally, Brockagh and Creevagh). Later work showed that some of these horizons were in fact folded repetitions of the same unit, (Rickard 1971; Hutton, 1977). The Creeslough Fm. grades by intercalation with graphitic pelite into the *Ards Pelite Fm.* In the Errigal area and NE of the Lackagh River the transition is marked by a limestone band known as the *Altan Limestone*, (Rickard, 1962; Hutton, 1977).

The Ards Pelite Fm.:- these pelitic rocks outcrop in a narrow strip from Fanad to Dunlewy, with isolated exposures occupying the cores of two anticlinal structures on the Crohy Hills, (SW of Dunglow). The Ards Pelite is a black, carbonaceous, pyritiferous pelite assemblage, containing occasional bands of impure limestone and impure quartzites. Towards the top of this package there is a rapid increase in psammitic units, getting progressively thicker upwards until there are only thin partings of pelite. The thickness of this distinctive "transition member", (McCall, 1954); Rickard, 1962) is variable, though on average is 30-60 metres.

The Ards Quartzite:- the most extensive outcrop of this unit is to the NW of the Main Donegal Granite, forming an area of shallow dipping beds between the Bloody Foreland, Crockator and Dunlewy. The prominent Errigal-Muckish range of mountains are also composed of this quartzite. More isolated exposures of the Ards Quartzite are seen at Horn Head, Rosguill, Fanad, Aranmore and Tory Islands, Crohy Hills, Gweebarra Estuary, Glenieragh, (NE end of Main Donegal Granite) and at

Lackagh Bridge. This formation is a thick sequence of homogeneous, clean, well-washed feldspathic sandstones, accompanied by bands or lenses of coarser grit, plus rare intercalations of non-graphitic pelite (rarely exceeding 1.0 metre in thickness). Bedding is indicated by grain size variations, feldspathic bands, plus thin muscovite bands and also by the presence of dark, dense mineral bands. Sedimentary structures are well preserved providing important way-up criteria in highly deformed areas, and include cross-bedding, graded bedding, ripple marks, washouts and small scale slump structures.

The Sessiagh-Clonmass Fm.:- the uniform nature of the Ards Quartzite is succeeded by this diverse assemblage of rocks consisting of alternating units of quartzite and limestone, (McCall, 1954) though lateral facies variants are seen in western Donegal, Iyengar *et al.* (1954) and Pitcher & Shackleton, (1966). The majority of the Imperial College Research team encountered rocks of this formation within their individual areas with exposure extending from Fanad to Dunlewy, (McCall 1954; Rickard 1962; Knill & Knill, 1961). On Breaghy Head, detailed structural mapping by Jolley (1996) has shown that the sequence of units proposed by McCall (1954), (see figure 2:3) are in fact tectonic duplications of the same units within a ductile imbricate stack. The Clonmass and Marble Hill Limestones and the Clonmass and Sessiagh Quartzites being the same units. The new sequence being Clonmass Limestone, Sessiagh Quartzite and Port Limestone, (Jolley 1996), the same sequence order visible on Rosguill, (Anderson 1978). The Clonmass Limestone is dolomitic in character having brown weathered surfaces and cream-pink fresh surfaces. This unit passes up into a transitional unit which has a calcareous base, though upwards these are replaced by calc-pelites, pelites, semi-pelites into interbedded quartzites and pelites. Above this transition member is the Sessiagh Quartzite which typically has fine-grained banded quartzites interbedded with more micaceous partings. This unit also contains inpersistent lenses of impure limestone. At the top of the sequence is the Port Limestone, a distinct blue schistose dolomite containing buff coloured dolomitic flags and micaceous partings, (McCall 1954). Along the south-eastern margin of the Main Donegal Granite there is an extensive strip of the Sessiagh-Clonmass Fm., (Pulvertaft 1961) which can be correlated through the raft-trains in the Main Donegal Granite at Glenleheen into the Maas area, Iyengar *et al.* (1954), (see figure 2:3) in western Donegal, where well differentiated sediments of the Creeslough-Errigal area are replaced at Maas by a thinner sequence of argillaceous limestones and calcareous mudrocks, (Pitcher & Berger 1972; Hutton & Alsop 1996), representing a deepening of the basin over this relatively short distance. The presence of the clearly identifiable, stratigraphically higher Lower Falcarragh pelites above the varying facies of this formation allows correlation between the respective areas, (Pitcher & Shackleton 1966).

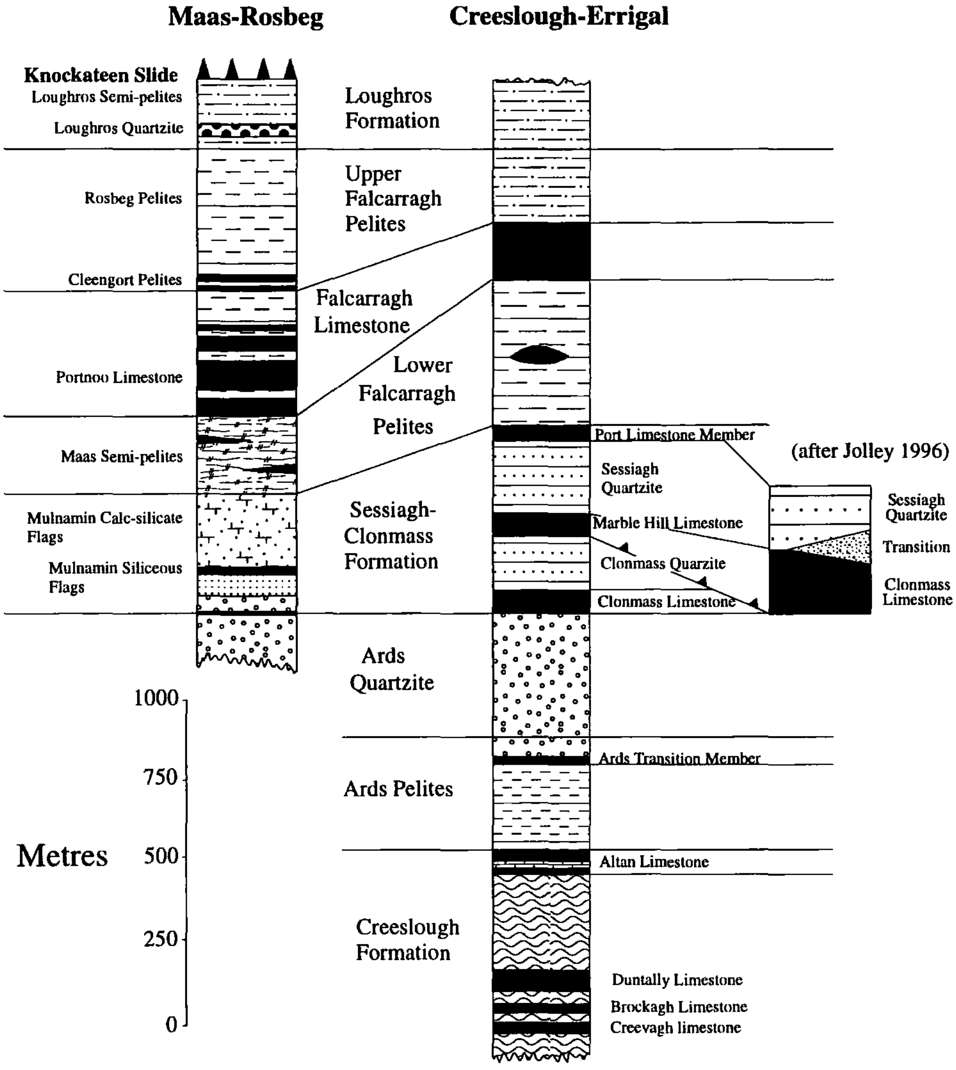


Figure 2:3:- facies variation within the Appin Group of NW Donegal, (modified from Pitcher & Berger 1972).

The Lower Falcarragh Pelites:- a thick pile of homogeneous pelites above the Sessiagh- Clonmass Formation, everywhere in Donegal. These pelites outcrop most extensively in the Falcarragh-Dunfanaghy area, with more isolated exposures on Rosguill and Fanad, though here they have been greatly disrupted by the intrusion of the Fanad Pluton. In these areas the Lower Falcarragh Pelites are typically up to 600 metres of grey and black striped, pyrite bearing, carbonaceous shales containing occasional bands of semi-pelite and calc-pelite. In the Gweebarra Estuary area, Iyengar *et al.* (1954) reported the presence of 200 metres of purplish, homogeneous pelites containing thick bands of more calcareous material. At Gartan Lough, on the

SE side of the Main Donegal Granite, the formation consists of alternating pelites and semi-pelites, though exposure is generally poor here.

The Falcarragh Limestone:- this limestone takes its name from the extensive outcrop of this unit in the district of Falcarragh and Gortahork, where Rickard (1961) observed it separating two thick units of pelite. The limestone has a bluish-grey appearance, containing rare intercalations of psammite and pelite, which become more numerous towards the top of the formation. Around the Gweebarra Estuary the limestone is very well exposed, and reaches a thickness of 225 metres. Pitcher & Shackleton (1966) traced it from here, through the Main Donegal Granite, onto the SE margin of the pluton where it forms two strips by a presumed folded repetition. The good exposure around Claggan Lough, (near Lough Gartan) and Lough Salt allowed Pitcher & Shackleton (1966) to correlate the rocks of this area (Fintown Succession) with the remainder of the Creeslough Succession.

The Upper Falcarragh Pelites:- the incoming of pelitic and psammitic layers to the top of the underlying limestone forms a clear transition between these two formations. In the type locality these pelites have a striped appearance due to the presence of lighter coloured semi-pelitic layers and the lower abundance of carbonaceous material when compared to the Lower Falcarragh Pelites. Similar pelites are well exposed at Maas, (the Cleengort pelites of Iyengar *et al.* (1954) and form much of the envelope to the Ardara Pluton. In this area the lower part of the formation is markedly pelitic, grading up into more semi-pelitic material, which contains a number of limestone bands. The more semi-pelitic character of the Upper Falcarragh Pelites in this area led Pitcher & Shackleton (1966) to assign the Rosbeg Semi-Pelites, (Akaad 1956) to this formation, as this feature is seen at the type locality. Two strips of Upper Falcarragh Pelite also form a folded repetition, extending from Fintown to Lough Greenan, (near Lough Salt), (Pitcher & Shackleton 1966).

The Loughros Formation:- forms the highest formation within the Creeslough Succession of Pitcher & Berger 1972), although its exposure is restricted to isolated areas. In all of these areas this formation follows the Upper Falcarragh Pelites, and is typified by a composite package of quartzites and semi-pelites. The most continuous exposure of this is seen on the Loughros Peninsula, extending NE towards Glenties and Fintown. On Inishbofin, NE of Bloody Foreland, Rickard (1962) observed similar rocks in contact with the Thorr Pluton. Finally, Knill & Knill (1961), found members of the Loughros Fm., on the northern part of Rosguill, as "ghost-stratigraphy" within the Fanad Pluton, although here calcareous flags, limestones and semi-pelites are more typical.

Generally the rocks of the Appin Group are laterally continuous and consist of well differentiated sediments which show limited facies change, (except the Sessiagh Clonmass Fm). The type of sedimentary structures strongly suggest that all the units were deposited in relatively shallow waters such as a shelf, though conditions may have either deepened or became more restricted allow the slow deposition of the Ards Black Pelite. The overlying quartzite clearly indicates a return to shallow water deposition, with the later development of lagoonal conditions during the deposition of the Sessiagh-Clonmass Fm., (e.g. mud cracks) with minor differential subsidence occurring causing localised facies variation. Regional subsidence continued allowing the deposition of the Falcarragh Pelites in a possible sub-tidal to offshore or basinal environment, with these conditions being maintained, but extending over wider area allowing the deposition of the remainder of the Appin Group, (Harris & Pitcher 1975).

2:3:2: The Argyll Group

These rocks originally called the Kilmacrenan Succession, Pitcher & Berger (1972) can be traced from NW Inishowen, through Fanad, central and southern Donegal, extending into the Slieve League Peninsula. For clarity it is suitable to describe this succession in two parts; the rocks to the NW of the Leannan Fault, and those to the SE of this prominent dislocation.
For correlation between the two main areas see figure 2:4.

2:3:2:1:-*The Argyll Group to the NW of the Leannan Fault*

The strike of these units is approximately NE-SW, running sub-parallel to the Leannan Fault for a distance of 48 km, (Pitcher *et al.* 1964) with the succession younging towards the SE. The standardised succession has been taken from (Pitcher & Berger 1972), i.e. the Kilmacrenan Succession, although for a recent comparison see figure 2:4.

At the base of the succession are the *Glencolumbkille Pelites* forming isolated exposures often disrupted by the Knockateen Slide. These pelites are dominantly graphitic with partings of psammite and graphitic limestones, and passes up everywhere into light grey marble, known as the *Glencolumbkille Limestone.* This unit consists of two limestone members which sandwich a grey striped pelitic member forming two main outcrops on Fanad and Glencolumbkille. The above two formations belong to the Appin Group. The *Port Askaig Tillite* is next marking the base of the Argyll Group in the Dalradian. The tillite is visible in both Scotland and Ireland, forming an important stratigraphic marker, although the thickness in Donegal is much less than that seen at the type locality on Islay. The clasts within the tillite vary in size (0.001-1.0 metre) and composition, though there are two main types.

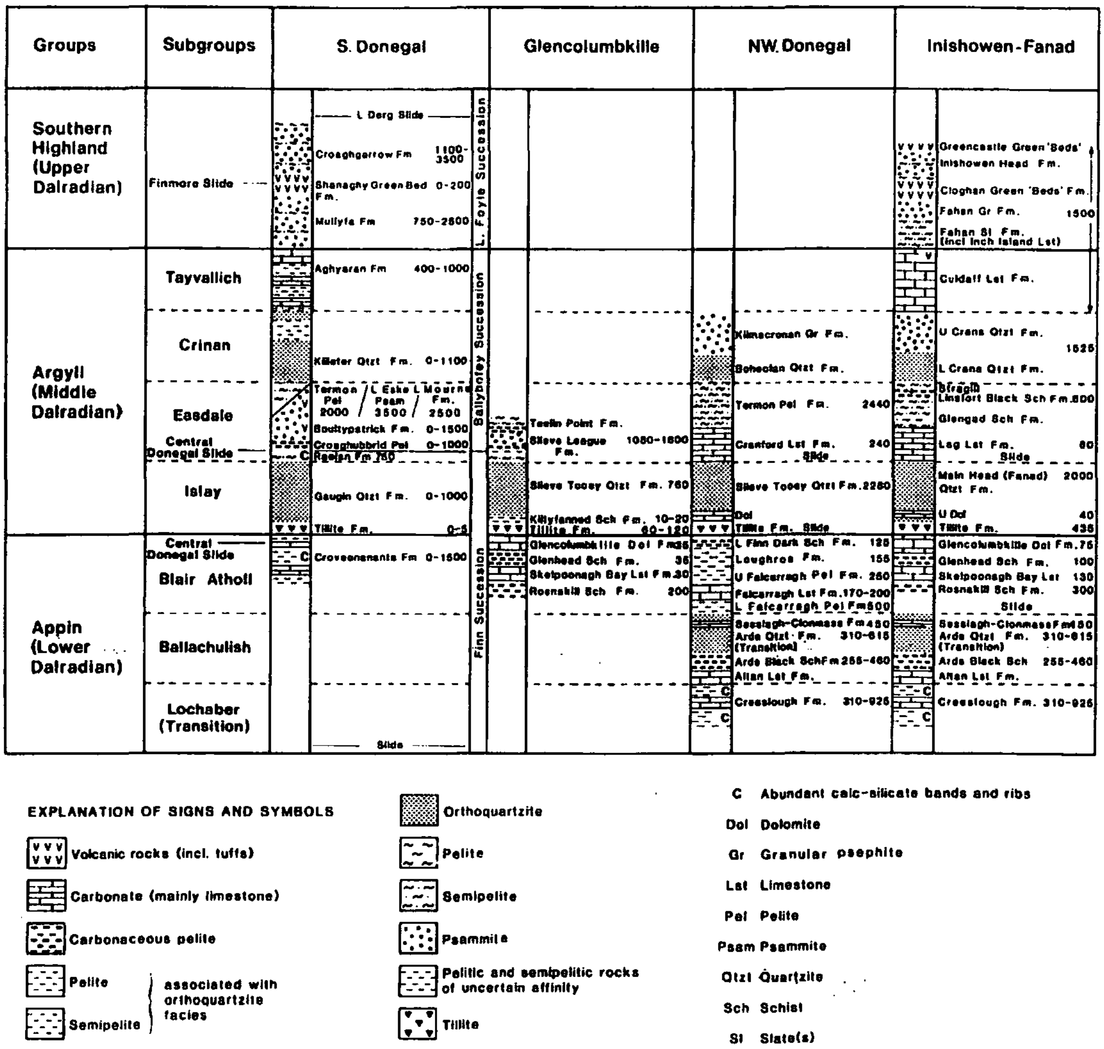


Figure 2:4:- table showing the correlation of the Dalradian Supergroup in Donegal, Ireland, (after Alsop & Hutton 1990).

Towards the base the clasts are mainly dolomitic and are believed to be of local derivation, (intra-basinal). At higher stratigraphical levels clasts composed of igneous and metamorphic fragments, e.g. gneiss, granite and quartzites become more common suggesting derivation from basement highs outside the basin, (extra-basinal).

Above the tillite is the *Slieve Tooey Quartzite*, a quartz arenite forming one of the thickest formations, (up to 2300m) within the Argyll Group. The most extensive exposure is on the northern part of the Slieve League Peninsula and the Aghla Mountains, (south of Fintown), whilst in the following 25 km to the NE much of the quartzite has been cut-out along the Knockateen Slide. In the Lough Salt area the exposure becomes substantial forming the contact to the Main Donegal Granite around the Glen area where it is juxtaposed against the Ards Quarzite. Further exposures extend into Fanad and Inishowen.

At the top of the Slieve Tooey Quartzite in NW Donegal (Inishowen-Glenties) there is a rapid change into calcareous rock types, known as the *Cranford Limestone*. This formation, at the type locality, consists of up to 240m of yellow weathering dolomite, which pass into grey limestone interbedded with pelite. Pitcher & Berger (1972) put the Cranford Limestone as the basal member of the overlying pelites-the Termon Fm. because at Glenties (and south-westwards to the Slieve League) the Slieve Tooey Quartzite is overlain by pelites. At Slieve League there is a thick sequence of dark pelites and psammites, (Kemp 1966) overlying the Slieve Tooey Quartzite which were grouped together as the Slieve League Formation, (Pitcher & Berger 1972). Thus, there is a major facies change from shallow water in the NE to more basinal facies to the SW. The Slieve League Fm. has been subsequently correlated with the lower part of the *Termon Pelite Formation*, (Alsop & Hutton 1990) (see figure 2:4). The Termon Pelites are a thick (up to 2500m), continuous formation in both Ireland and Scotland. In NW Donegal the formation consists of a lower dark, graphitic pelites containing occasional lenses of dolomite and arenite, which grade upwards into calcareous, chloritic pelite and fine-grained green psammites (Pitcher & Berger 1972). Towards the top of the Termon Fm. the pelites become darker again, with the incoming of psammite intercalations which thicken upwards into the *Crana Quartzite*. To the SE of the Leannan Fault the Crana Quartzite is divided into three units, (McCallien 1937), whilst to the NW of the fault the lower members have been cut-out by movement along the Leannan fault complex. The "Lower Crana Quartzite" is thick units of fine psammite interbedded with striped semi-pelites, passing up into striped psammites interbanded with thin limestones, (The "Killygarvan Flags"). The "Upper Crana Quartzite" sees a return to a package of graded pebbly grits. The three divisions which comprise the Crana Quartzite are interpreted as being turbiditic in character, (Pitcher & Berger 1972).

Above the Crana Quartzite there is a sharp change into thick limestone interbedded with graphitic pelites, the *Culdaff Limestone*, (McCallien 1935), which is only exposed in Inishowen. The Culdaff Limestone has been correlated with the Loch Tay Limestone in Scotland, the top of which represents the junction between the Argyll and Southern Highlands Groups, (Pitcher *et al.* 1970; Pitcher & Berger 1972; Harris & Pitcher 1975).

2:3:2:2:-The Argyll Group to the SE of the Leannan Fault

The Appin Group of central Donegal, to the SE of the Leannan Fault, records a major lateral facies change. In Inishowen, (on the SE side of the Leannan Fault) the more typical rocks of the Argyll Group previously described rapidly change into rocks which have affinities with deep water basin sedimentation. The presence of certain marker horizons within the Argyll Group allows correlation between the two areas, (Pitcher *et al.* 1971; Alsop & Hutton 1990).

Earlier workers, (Pitcher *et al.* 1964) initially thought the rocks of Central Donegal became older towards the south until eventually they unconformably lay on Pre-Caledonian basement, (Lough Derg Psammites) which was believed to belong to the Moine Series, (Anderson 1947). The work of Pitcher *et al.* (1970), discovered a kilometre scale fold they called Ballybofey Anticline which inverted the rocks to the south of the axial trace, i.e. the Dalradian rocks of southern Donegal, therefore young towards the south until rocks of the Lough Foyle Succession are juxtaposed against the Lough Derg Psammites, separated by the Lough Derg Slide, (Wood 1970).

Alsop & Hutton (1990), modified the stratigraphic nomenclature of central Donegal by erecting two tectono-stratigraphic successions that are separated by the Central Donegal Slide, (see figure 2:5). The rocks consistently young away from this structure and belong to either the ***Finn*** or ***Ballybofey*** Successions, with the former containing the older rocks. For a correlation of these two tectono-stratigraphic successions with the rocks to the NW of the Leannan Fault (see figure 2:4).

The palaeogeographic reconstruction of the Argyll Group implies that after the Varangian Glaciation event there were large volumes of clastic material available resulting in the deposition of the thick, shallow-water quartzites with cross bedding, mud cracks and extensive reworking. The nature of the sediment implies erosion of granitic gneissic basement, similar to the extra-basinal clasts seen in the upper parts of the tillite, (Harris & Pitcher 1975) Above Slieve Tooey Qtz. there is a considerable facies change with the shallow water Cranford Limestone, (basin margin platform facies of Alsop & Hutton 1990) giving way south-westwards to thick pelite and psammite deposits of the Slieve League Formation. Some workers have suggested axial derivation of these sediments along a NE-SW trough, (Harris & Pitcher 1975). A similar facies change is in the Termon Pelite Fm./Lough Eske Psammite Fm./Lough Mourne Fm. with increasing water depths and coarsening of sediments occurring to the south which may be related to the development of sub-basins in southern Donegal. Alsop & Hutton (1990) report green beds in the Termon Pelite Fm., suggesting volcanics in response to crustal extension. The rocks towards the top of the Argyll Group show turbiditic affinities, a style of deposition continuing into the overlying Southern Highlands Group.

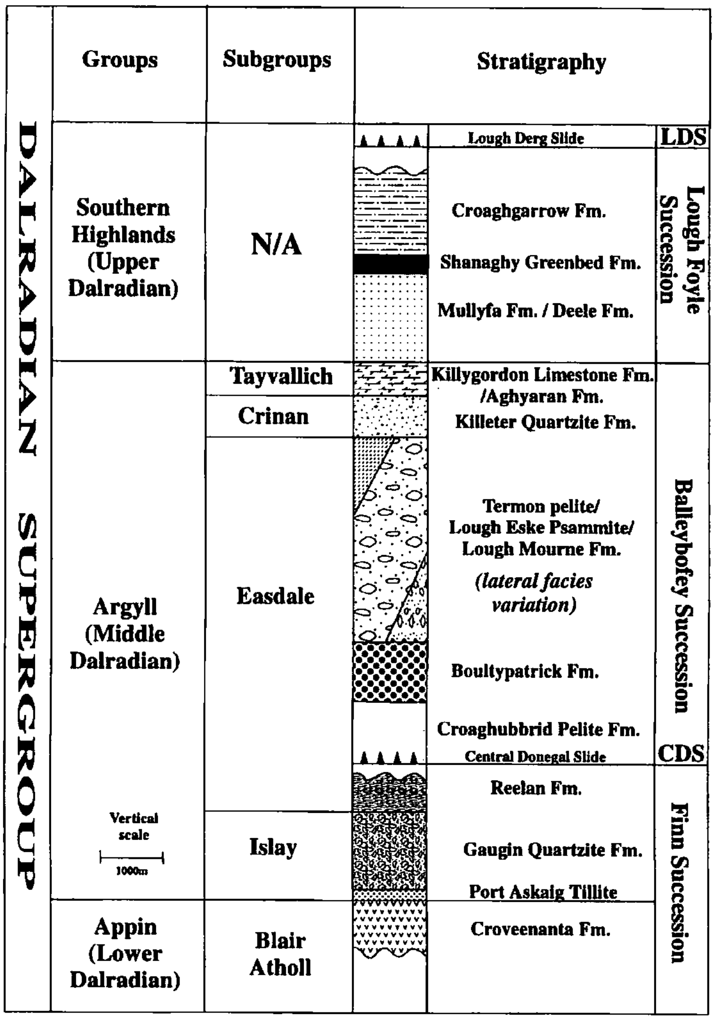


Figure 2:5:- the Argyll and Southern Highland Groups of central and southern Donegal, (*after* Alsop & Hutton 1990).

2:3:3: The Southern Highlands Group (the Lough Foyle Succession)

In contrast to the rocks of the Appin and Argyll rocks in Donegal, the Lough Foyle Succession, (Southern Highlands Group) consists of relatively monotonous sequence of turbiditic rocks which outcrop in South Donegal, the Sperrin Mountains, (Omagh and Tyrone) and much of land bordering Lough Foyle. Within the Lough Foyle Succession there are volcanogenic green beds, (Harris & Pitcher 1975), suggesting increased tectonic instability within the Dalradian Basin. Throughout the Dalradian rocks of Donegal large volumes of tholeiitic basic magma have been emplaced and will be discussed in the following section.

2:4: The meta-basites of Donegal

Within the Dalradian sediments of Donegal there is a series of intrusive tholeiitic sills, (now metadolerite) of varying dimensions, which have witnessed varying degrees of deformation and metamorphism associated with regional events and by the later emplacement of granite bodies. Two types of dolerite have been observed in Donegal, (Pitcher & Berger 1972) the more common type is a *quartz dolerite* which produces static contact hornfelses within the host sediments. The other type of dolerite forms a suite on the Rosbeg Peninsula, (*the dolerites of Rosbeg*) and are characterised by coarse-grained garnet amphibolites, (intruded into phyllitic schists of the Upper Falcarragh Pelites), that bleach the immediate host forming a white adinole which grades into normal pelite, (usually over distance of 3-5 metres). All primary features of the Rosbeg dolerites have been obliterated by subsequent deformation resulting in schistose amphibolites. The intrusion of the Rosbeg dolerites appears to predate the earliest deformation, (Obaid 1967).

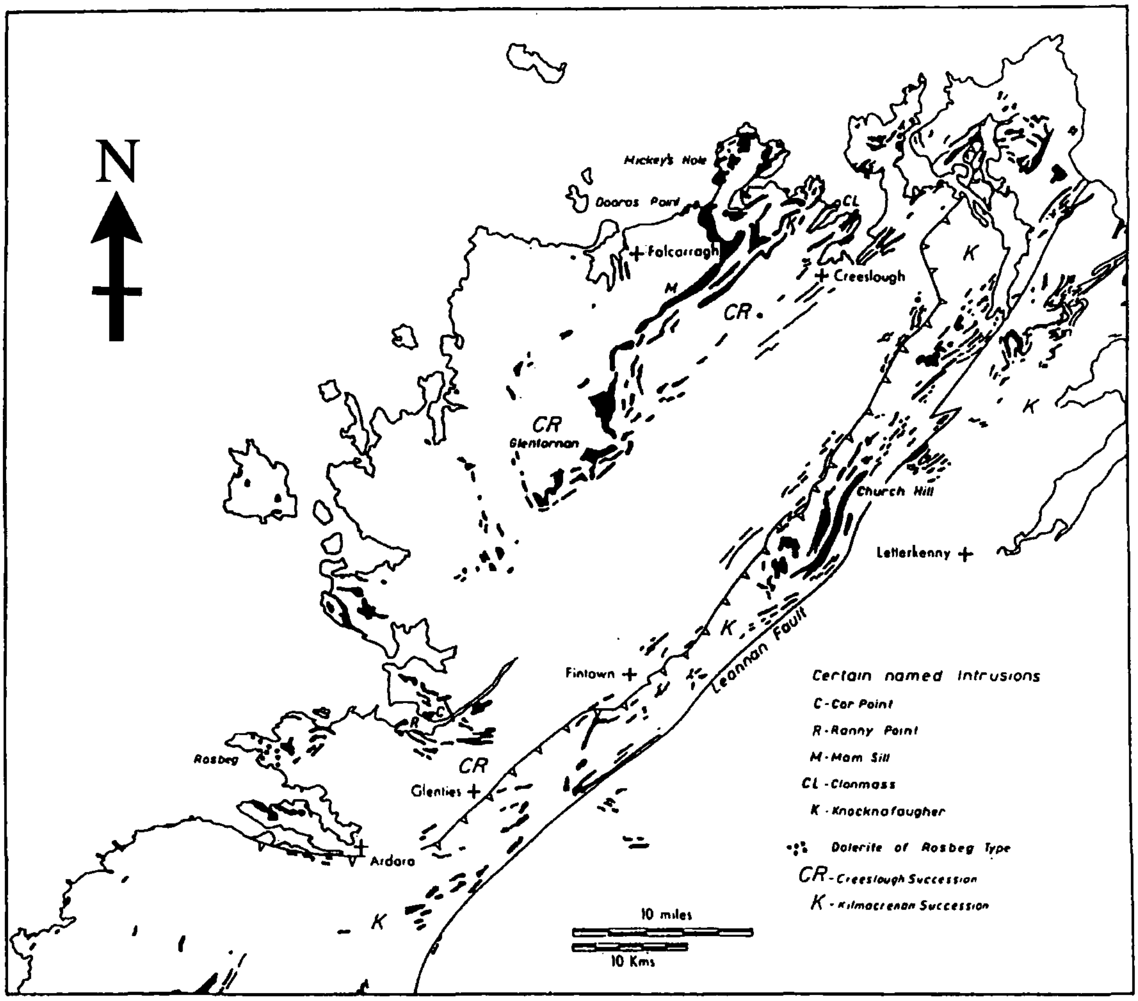


Figure 2:6:- map illustrating the distribution of metadolerite intrusions in NW Donegal, (after Pitcher & Berger 1972).

The quartz dolerites are volumetrically more important forming a plexus of sheets in the Dalradian of Donegal. The sills are mostly concentrated in the Creeslough-Errigal area and the Kilmacrenan-Milford area, (see figure 2:6) where they locally comprise up to 25% of the rock pile, (Hutton & Alsop, 1996). In central Donegal the dolerites are far less common. Hutton & Alsop (1996) have related this distribution to the siting along a major NNE trending lineament. The dolerites mostly form sills or slightly transgressive sheets, varying in thickness from 0.5 to 200 metres. They are often traceable for many kilometres, forming useful marker horizons, e.g. the Mam Sill, Ranny Point Sill, Knocknafaugher Sill and Clonmass Sill, (figure 2:6 for location). In the centre of the thickest sills, which resisted the deformation and metamorphism, primary ophitic textures have been preserved, although the original minerals are only rarely preserved, e.g. on Rosguill, (Pitcher & Berger 1972). Original minerals were plagioclase, (labradorite-andesine), augite, quartz and ilmenite with alteration to hornblende, albite, clinozoisite, quartz and ilmenite, depending on the intensity of deformation or vicinity of granitic intrusions, (or both).

Where dolerites have intruded into pelites and semi-pelites they have produced narrow static aureoles which have protected these rocks from subsequent deformation and metamorphism, preserving original sedimentary features, (Pitcher & Berger 1972).

The emplacement of the dolerites have been treated, by most early workers, as been pre-deformational, although some workers, (Pitcher & Berger 1972) and (Hutton 1979) believe some of the quartz dolerites were emplaced during the deformation. Hutton (1979) reported at Dunlewy that S_2 foliations are hornfelsed by the 200 metre thick Mam Sill, clearly implying this dolerite unit is later, a possibility first suggested by Rickard (1961). In broad terms, Hutton (1979), believes the majority of the quartz dolerite was emplaced after the major folding, (D_2) but prior to D_3 deformation, (see following sections) also correlating with the peak of regional metamorphism. The implication, therefore, is that the dolerites belong to several suites and are not all the same age, with the emplacement of the later sills occurring during crustal thickening rather than crustal thinning (Hutton 1979).

2:5: The pre-granite Caledonian structure of the NW Irish Dalradian

The Dalradian rocks of Donegal have been subjected to a protracted history of deformation associated with regional Caledonian tectonics, termed by some workers the Grampian Orogeny, (Lambert & Mckerrow 1976; *in* Hutton 1983). The architecture of these early structures have possibly influenced the siting and form of the plutons which comprise the Donegal Batholith. Later, more localised deformation, (approx. 400 Ma) is associated with the intrusion of the Donegal

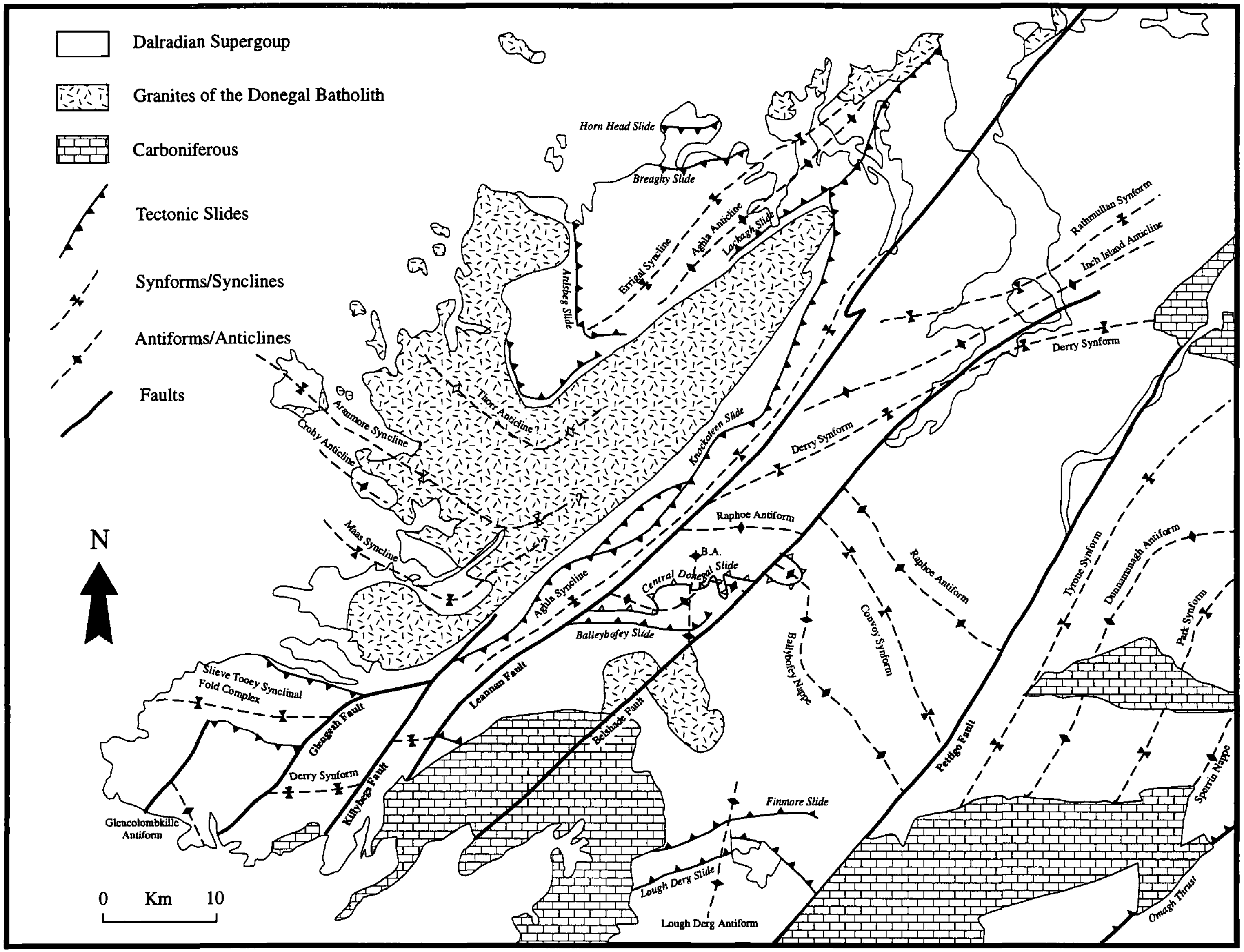


Figure 2:7:- Structural map of NW Ireland showing the traces of major folds, faults and tectonic slides, (compiled from Pitcher & Berger (1972) and Hutton & Alsop (1996))

Granites, most notably the Main Donegal Granite and its intensely deformed aureole relating to a sinistral transcurrent shear zone, (Hutton 1982). During the Devonian, the basement rocks and granites of Donegal were cut by a whole series of NE-SW trending faults with predominantly strike-slip movements to them, e.g. the Leannan Fault, (Pitcher *et al.* 1964; Alsop 1992a) and the Belshade Fault, (Walker & Leedal 1954), (see figures 2:2 and 2:7).

2:5:1 The structural history of the Appin Group

The structural evolution of the Creeslough Succession will be discussed in detail first, later moving onto the rocks of central and southern Donegal to obtain a regional picture of the deformation suffered in the Irish Caledonides.

2:5:1:1 NW Donegal (north of Main Donegal Granite)

Although many major structures were recognised by an earlier generation of workers, e.g. Rickard, McCall, Pulvertaft etc., the first complete tectonic synthesis of the Appin Group of NW Donegal did not appear until Pitcher & Berger (1972) after the pooling together of previous data plus the new work of their own, e.g. Berger & Cambray. The area to the NW of the Main Donegal Granite will be addressed first, due to it being the largest area of exposure. The chronology and style of deformation will then be compared and contrasted with the other areas of the Appin Group in Donegal.

The structure of this area is dominated by a series of major NW facing, gently inclined recumbent folds and associated tectonic slides, (Hutton 1983) accompanied by a pervasive axial planar cleavage which is prominent throughout the area, (assigned to D_2 by Pitcher & Berger 1972; Hutton 1977). Later deformation intensifies south-eastwards, becoming the dominant tectonic fabric towards the margin of the Main Donegal Granite. This is related to more localised movement along the Main Donegal Granite Shear Zone (MDGSZ), which later influenced the emplacement of this pluton (Hutton 1982).

The chronology of deformation will be treated in terms of "D", "F" and "S" numbers, as used by earlier workers.

D_1: within NW Donegal no major F1 folds are observed, though S_1 forms a schistosity often parallel to bedding, (the "bedding schistosity" of Pitcher & Berger 1972). Early workers, (Rickard 1962) did not believe S_1 was of tectonic origin, inferring the planar mineral alignment was due to lithostatic loading during the early stages of metamorphism and folding. However, in certain locations, (Ards Priory and Rossapenna) often in the hinges of later folds where strains are lower, obliquity between bedding and S_1 can be seen, supporting a tectonic origin, with bedding facing up to the NW in S_1 (Hutton 1977; Anderton 1978). D_1 is associated with lower greenschist facies metamorphism, with a preferred orientation of chlorite, muscovite and quartz.

D_2:- the major folds and associated ductile tectonic slides have been assigned to D_2 in NW Donegal, (Hutton 1977; 1983). The peak of regional metamorphism was syn- to late D_2, (Hutton 1983; Meneilly 1982) at mid-upper greenschist facies. The most prominent fold is the *Errigal Syncline*, (Rickard 1962; Pitcher & Berger 1972) first identified by McCall (1954) at Marble Hill Strand, and later traced into the Errigal area by Rickard (1962), who also identified the complementary and overlying *Aghla Anticline.* A series of more minor F_2 folds were found to the NW on Horn Head; the Polnaguill Syncline and Dooros Anticline, (McCall 1954; Pitcher & Berger 1972). All of these folds face towards the NW and have gently to inclined axial planes, dipping to the SE. S_2, (composite $S2/S_3$ of Pitcher & Berger 1972) is axial planar to the major folds and can be observed to change vergence across the Errigal Syncline, (where it cross-cuts bedding it verges NW in right-way-up rocks and verges S to SE in inverted rocks), indicating that S_2 is axial planar to the fold and hence it is of F_2 age.

According to Hutton (1983) the F_2 folds are synchronous with tectonic sliding, unlike the views of previous workers, (McCall 1954; Knill & Knill 1961; Rickard 1961; Pitcher & Berger 1972) who believed the slides to be D_1 in origin. McCall (1954), originally proposed that the prominent Horn Head Slide had "lagged"* the succession above the Ards Quartzite to the NW, on the normal limb of the Errigal Syncline, (see figure 2:8a). The identification of two slide planes at Horn Head led McCall (1954) to believe that they were the folded repetition of the same feature, bent around the so-called Polnaguill-Dooros fold pair. On Rosguill, Knill & Knill (1961), also reported the folding of the Horn Head Slide around the continuation of the Errigal Syncline into this area. The apparent folding of the slide led these workers and the later work of Pitcher & Berger (1972) to put a D_1 age for the slides as they were clearly folded around F_2 folds. Work on Horn Head by Hutton (1983) showed that the Horn Head Slide was in fact two separate slides, (the Horn Head and Breaghy Head-Dunfanaghy Slides), which both have thrust geometry, transporting the hanging-wall to the NW, (see figure 2:8b). The folding of the Horn Head Slide around the Errigal Syncline at the top of the Ards Quartzite was disproven at Marble Hill, (Hutton 1977) and on Rosguill, (Anderson 1978). The Polnaguill-Dooros fold pair are thus regarded as not existing and are replaced with the F_2 Horn Head Anticline, Muntermellan Monocline, Figart Syncline and Dunfanaghy Anticline associated with the Horn Head and Breaghy Head Slides respectively, (Hutton 1983). The structure above the

* "lagging":- a situation arising where younger beds are tectonically placed, by a fault or thrust, onto younger stratigraphic units.

Breaghy Head Slide is interpreted as a large dislocated sheath fold produced by high shear strains associated with sliding causing rotation of fold axes towards the transport direction, (Hutton, 1983).

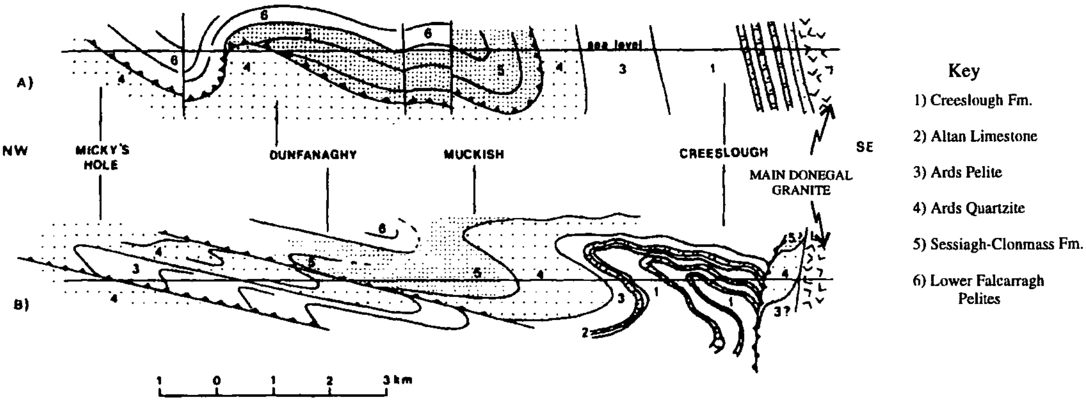
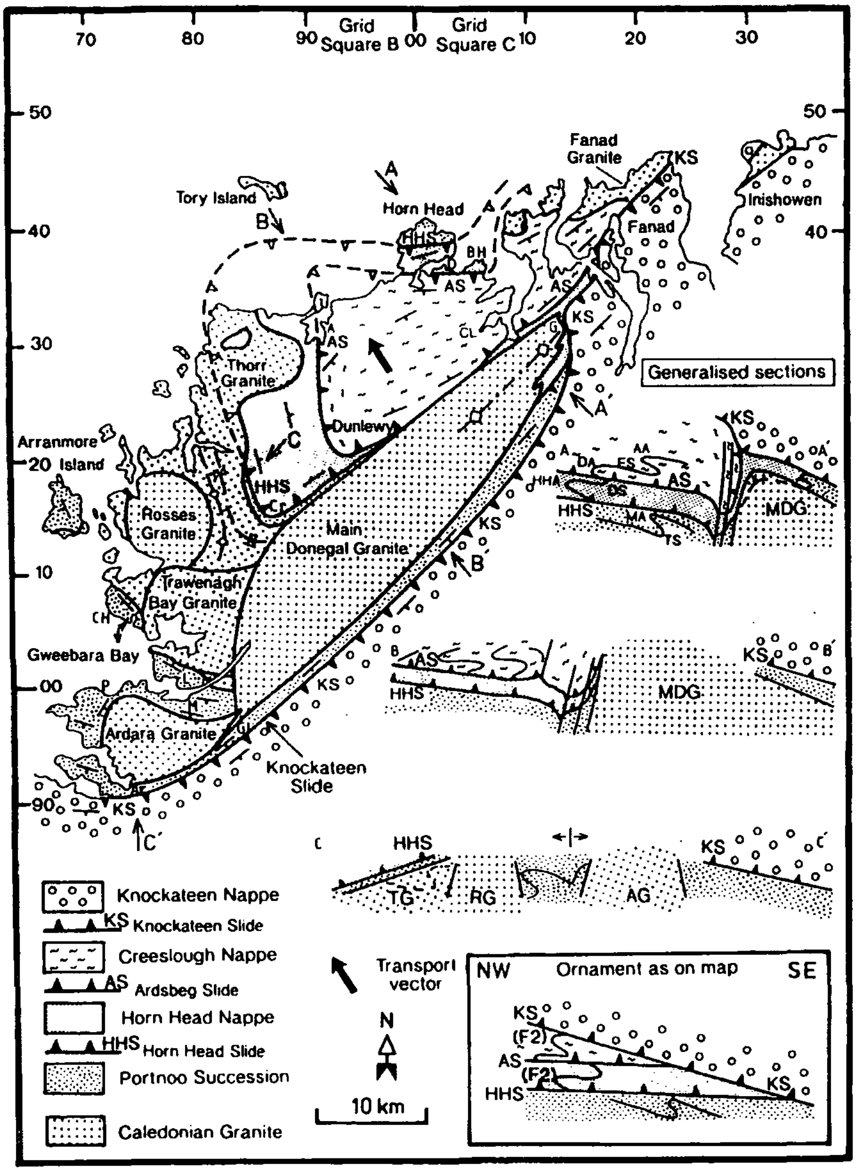


Figure 2:8:- sketch cross-sections through the Appin Group to the north of the Main Donegal Granite, in the Creeslough area:- (a) after McCall (1954). (b) after Hutton (1977).

The closure of this fold to the NW also indicates the sense of shear, i.e. towards the NW. Recent work by Jolley (1996) on Breaghy Head has shown that this slide has at least 13 ductile imbricates which have a cumulate displacement of some 6 km towards the NW. D_2 in this area is a diachronous event with Jolley (1996) reporting the folding of the detachment and the structurally highest imbricate folded around the Errigal Syncline. Hutton & Alsop (1995) have correlated the Horn Head Slide as being the same as the slide seen at Crockator mountain, (Rickard 1963), which has Upper Falcarragh pelite in the footwall, with a shallowly dipping sheet of Ards Quartzite above. The same authors correlate the Breaghy Head Slide with the Dunlewy-Ardsbeg Slide, which lies above the Horn Head Slide.

D_2 deformation has been related to the emplacement of foreland propagating nappes carried north-westwards along a series of ductile slides, (Hutton & Alsop 1995). The structurally lower Horn Head-Crockator Slide carries the Horn Head nappe, containing the F_2 Horn Head Antiform. Below the Horn Head Slide is the Portnoo Succession, (Hutton & Alsop 1995) which also contains NW facing F_2 folds, inferring the presence of a thrust at deeper structural levels. Above the Horn Head nappe is the Ardsbeg Slide, (correlated with the Breaghy Slide) which carries the overlying Creeslough nappe. This nappe contains the Errigal Syncline-Aghla Anticline fold pair, plus a high degree of imbrication associated with ductile thrusts, (Jolley 1996). The Creeslough nappe is overridden by the Knockateen Slide carrying Argyll rocks over the Appin Group. The presence of Portnoo succession rocks in the footwall of the Knockateen Slide implies that both the Horn Head Slide and Ardsbeg Slide have been truncated (see figure 2:9). Alsop & Hutton (1995) estimate values of displacement of 100-250 km.



Abbreviations

KS - Knockateen Slide
AS - Ardsbeg Slide
HHS - Horn Head Slide

AA - Aghla Anticline
ES - Errigal Syncline
DA - DunfanaghyAnticline
DS - Dunfanaghy Syncline
HHA - Horn Head Anticline
MA - Maghery Anticline
TS - Thorr Syncline

Sections

MDG - Main Donegal Granite

TG - Thorr Granite

RG - Rosses Granite

AG - Ardara Granite

Locations

D - Dunfanaghy
BH - Breaghy Head
CL - Creeslough
A - Ardsbeg
G - Glen
Cr - Crockator
CH - Crohy Hills
L - Lettermacaward
M - Maas
P - Portnoo
GL - Glenties
Ar - Ardara

Figure 2:9:- map of NW Donegal showing a nappe interpretation of the major D_2 structures, (after Hutton & Alsop 1995).

D_3:- Pitcher & Berger (1972) believed this to be a composite structure, (D_2/D_3) closely related to D_2, with the S_2 cleavage accentuated by a coaxial and coplanar deformation, (D_3) which they believed to be genetically related. Hutton (1977), however, separated these structures into two phases of deformation. S_3 **always** verges to the SE on S_2, with bedding facing up to the SE in S_3, (F3 axial planes) on the uninverted limbs of F_2 folds and faces down to the NW on F_2 inverted limbs, (Hutton 1977; 1983).

Therefore S_3 is clearly superimposed on S_2 with minor structures of F_2 and F_3 having opposing vergence and facing on the normal limbs of F_2 folds. S_3 usually dips gently to moderately towards the NW and has an obvious crenulating character in both outcrop and thin section. Larger D_3 structures include the Templebragga Antiform-Rough Point Synform fold pair which warps the Horn Head Slide, (Hutton 1983).

The relationship and geometry of D_1, D_2 and D_3 in the Creeslough area can be correlated with similar structures mapped NE, approximately along strike on Rosguill, (Anderson 1978) and on West Fanad, (White & Hutton 1985). Despite this correlation it is apparent that at least locally extensional crenulation cleavages genetically related to S_2 in D_2 mylonite zones, (Jolley 1996) have been confused in the past with the S_3 of Hutton (1983), (Hutton, *pers comm.*)

D_4:- is characterised by N-S folds, which have a general upright attitude with westerly vergence, and an axial planar cleavage crenulating previous fabrics. On Horn Head mesoscopic F_4 folds interfere with oppositely verging F_3 folds to produce box folds which locally affects the outcrop pattern of the Horn Head Slide, (Hutton 1983). Reports of F_4 folds are also seen at Breaghy Head and on Rosguill, (Hutton 1977; Anderton 1978).

D_5:- broadly coaxial with F_4 folds, although verging in the opposite direction, i.e. towards the east. The occurrence and intensity of D_5 is much more sporadic in distribution when compared to that of F_4.

Rickard (1962) identified one of these F_4 or F_5 sets of folds, (the Cashel Belt of Superimposed Folds) to be the result of the emplacement of the Thorr Pluton, whilst Hutton (1982) relates these F_4 and F_5 phases to the intrusion of the Thorr and Ardara plutons.

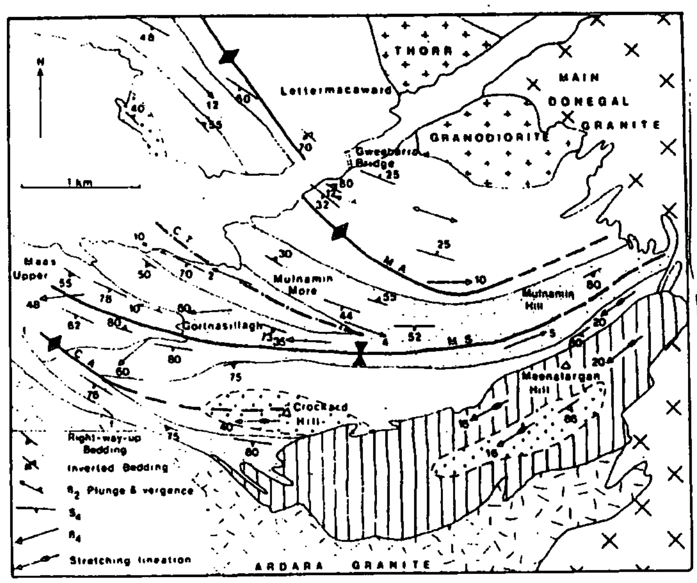
2:5:1:2 The Structure of the Gweebarra Bay area

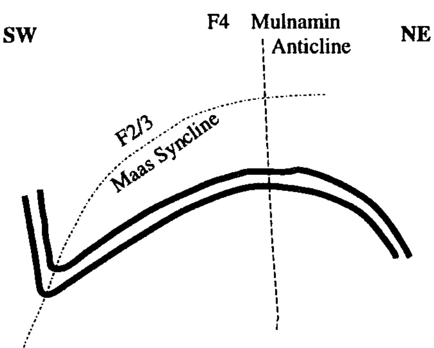
The rocks of the Gweebarra Bay are located on the western side of a major strike swing which results in a change in orientation of fold axes from NE-SW in N and E Donegal to NW-SE in W Donegal, (Meneilly 1982), (see later for further discussion). Although these areas were mapped by Gindy (1953), Iyengar *et al.* (1954); and Akaad (1956), little attempt was made to describe the polyphase deformation until the work of Meneilly (1982) from which much of the following account is taken. No major D_1 structures are recognised in the area although Hutton & Alsop (1996) report S_1 fabrics cross-cutting bedding in the Dawros-Loughros area. In general S_1 is transposed by the pervasive S_2. D_2 forms prominent folds in the area, i.e. on the Crohy Hills and at Rosbeg. S_2 is sub-vertical or dips south, becoming less steep to the north and progressively more parallel to bedding. Meneilly (1982) reports neutrally verging, upward facing F_2 folds in the south which pass northwards into northerly verging, NW facing folds. The peak metamorphic grade, indicated by garnet porphyroblasts, (upper-greenschist facies) occurred during D_2 and D_3. S_3 dips gently to the SW throughout the area and consistently verges to the SE on S_2. The orientation and geometry of D_2 and D_3 in this area, enabled Meneilly (1982) to correlate these structures with those observed in the Creeslough area, (Hutton 1977; 1983).

Meneilly (1982) divided the area into structural domains depending on the orientations of bedding, S_2 and S_3. Around Rosbeg, within the Upper Falcarragh Pelites, S_2 is at high angles to S_3 and both are at high angles to bedding; resulting in the development of major open F_2 folds with amplitudes of approximately 100-200 metres. Moving northwards into the Portnoo area, and extending eastwards into the Maas and Lettermacaward areas, S_2, S_3 and bedding progressively become into sub-parallelism, with major D_2 folds uncommon and later fold phases predominating. In the Crohy Hills, on the north side of the Gweebarra Estuary, Pitcher & Berger (1972) identified two $F_{2\text{-}3}$ structures one of which is more upright in the SE (Crohy anticline) whilst the second, (the Maghery anticline) becomes progressively overturned to be almost recumbent in the NW. Meneilly (1982) disagreed with this interpretation suggesting that this was one F_2 fold, (the Crohy Anticline) repeated by a fault.

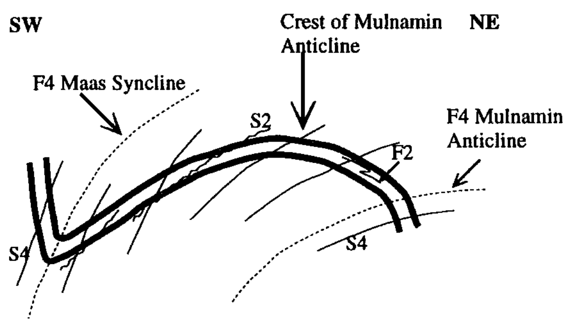
Structures relating to D_4 are more intense in Gweebarra estuary area, (figure 2:10a) compared to NW of the Main Donegal Granite, and major F_4 folds are found. Pitcher & Berger (1972) originally proposed that the F_2 Maas Syncline was folded by the upright F_4 Mulnamin anticline, (figure 2:10b). Remapping by Meneilly (1982) showed that the two structures were coeval with each other, finding evidence that the F_4 axial trace of Pitcher & Berger (1972) was in fact the crestal trace with the axial trace lying further to the north (figure 2:10c). Meneilly (1982) stated that D_4 is a regional deformation throughout the Gweebarra Bay area, though it intensifies rapidly towards the Ardara pluton and is generally related to the emplacement of this pluton. Thermal metamorphism in the aureole confirms the Ardara pluton is syn-tectonic with respect to F_4 folding. This is indicated by the growth of andalusite and staurolite porphyroblasts which overprint earlier S_4 cleavages but are tightened and augened by continued development of S_4. The presence of later biotite porphyroblasts associated with the Main Donegal Granite which overprint S_4 implies that the intensification of S_4 around porphyroblasts was related to the emplacement of the earlier Ardara pluton, (Pitcher & Berger 1972) and not coaxial flattening by the later Main Donegal Granite.

a) Structural map of the Gweebarra Bay area, NW Donegal, (*after* Meneilly (1982)





b) Pitcher & Berger, (1972) structural interpretation of the Maas area



c) Meneilly, (1982) re-interpretation of the Maas area

Figure 2:10:- Structures within the Gweebarra Bay area

2:5:1:3 The structure along the SE margin of the Main Donegal Granite

The structure of this area has not been reviewed in detail since the work of Pitcher & Berger (1972) from which the following account is taken. Earlier workers, Pitcher & Read (1959); Pulvertaft (1961) regarded this area as a right-way-up succession dipping steeply towards the SE around Fintown and becoming progressively more shallower towards the NE, around the Lough Greenan area. Pitcher & Shackleton (1966) showed there was repetition in strata and discovered major folds overturned to the NW, (the Lough Gartan Anticline and the Lough Greenan Syncline). Pitcher & Berger (1972), were uncertain about the age of these folds, (i.e. F_2/F_3 or F_4 the latter analogous to their Mulnamin Anticline). Around Lough Greenan, Berger (1967) observed gentle open, NE-SW trending folds, (his F_{4b}) which deform a recumbent anticline, a possible complementary fold to the Lough Greenan Syncline. Pitcher & Berger (1972) suggested the possibility of an F_1

structure in this area due to the outcrop patterns of metadolerites within the Falcarragh Limestone

On the Loughros Peninsula to the south of the Ardara granite, approximately along strike from the rocks mentioned in the above section, Meneilly (1982), observed that the style of folding was similar to that seen at Rosbeg, with F_2 facing towards the north and S_3 verging and facing to the south on normal limbs of F_2.

2:5:1:4 Deformation associated with the Main Donegal Granite

The intensely deformed aureole of the Main Donegal has been studied by many workers, (Pitcher & Read 1960; Pitcher & Berger 1972; Chenevix-Trench 1975; Hutton 1977; 1982; Hutton & Alsop 1995). In summary the deformation within the country rock intensifies towards the contact of the pluton, progressively obliterating structures related to earlier deformational events. The structures produced by this later deformation will be dealt with more fully when the Main Donegal Granite is discussed in the next chapter. For the sake of completeness the geometry of the structures associated with the pluton will briefly be described, allowing the construction of the deformational chronology for the Creeslough Succession, (Hutton 1982; 1983)

A detailed analysis of the structure of the aureole was performed by Hutton (1977) in a cross-section from Ards Priory to Lackagh Bridge. The D_{1-3} regional structures already described are identifiable in this area, but are overprinted by later structures. (**N.B.** the chronology described here will be that of Hutton (1983) and not Hutton (1977) in an attempt to avoid confusion with other areas, (see figure 2:11).

Structures relating to the D_4 and D_5, (Hutton 1983) events were not seen in the Lackagh Bridge area by Hutton (1977).

D_6: (D_4 of Hutton 1977). Away from the granite, F_6 folds trend ESE-WNW, progressively rotating in orientation towards the granite where they trend ENE-WSW and are typified by gentle to close apical angle and vergence approximately to the north. S_6 is axial planar to F_6 structures and becomes an intense, sub-vertical penetrative fabric with an associated sub-horizontal stretching lineation towards the granite. D6 produces large structures within the adjacent aureole, (see later).

D_7:- (D_5 of Hutton 1977). This is a comparatively weak event, S_7, easily confused with S_6; although the former dips at a moderate angle and can be observed to crenulate the S_6 foliation. D_7 produces rare open warps and small buckles.

D_8:- (D_6 of Hutton 1977) a crenulation cleavage of 'normal' type, (Cosgrove 1976) trending NNE-SSW, showing sinistral offset. Developed to a lesser extent is a conjugate crenulation cleavage, trending approximately E-W and displaying dextral

Rickard (1962 & 1963)	Pitcher & Berger (1972)	Chenevix-Trench (1975)	Hutton (1977)	White & Hutton (1985)	Meneilly (1982)	Hutton (1982 & 1983) REGIONAL
Bedding-schistosity	D1, S1 (Bedding-schistosity) Tectonic Sliding	Major recumbent isoclinal folds-tectonic sliding, D1, S1 & F1	D1, S1 & F1?	D1 & S1	D1 & S1	D1, S1 and F1?
Major folds, tectonic sliding and development of crenulation cleavage.	D2-D3 Composite structures { F2, S2 (major folds) / S3 (minor F3) }	D2, S2 & F2 Outcrop scale F2 folds Crenulating S1	D2:- Formation of major recumbent folds	D2 (major folds)	D2 S2 F2 (major folds)	D2 Formation of major recumbent folds. Flat lying S2
		D3, S3 & F3.:- formation of major, recumbent folds (e.g. Errigal Syncline)				
Dunlewy "Cross folds"		D3a:- sporadic minor F3 folds. S3 crenulation of S2., No major structures	D3:- formation of minor folds with opposite vergence	D3 folds verging opposite to F2	D3, S3 & F3 folds verging opposite to f2 folds	D3:- formation of minor folds with opposite vergence to F2.
The *Cashel Belt* of Superimposed folds (Related to the intrusion of the Thorr Pluton)	S4a:- crenulation cleavage with no major structures associated. F4b:- open upright folds with no associated minor structures. Relationship between S4a and F4b uncertain	D4:- F4 monoformal flexures. Weak crenulation	-	-	-	D4:- upright folds and cleavage. Folds trend N-S. Associated with the intrusion of the Thorr Pluton.
		D5:- open, upright NE-SW folds with steeply dipping S5 crenulation cleavage				
		D5a:- open upright folds F5a trending N-S with weak creulation cleavage	-	-	D4:- folds associated in time with the intrusion of the Ardara Pluton.	D5:- upright folds and cleavage. Folds trend N-S. Associated with the intrusion of Ardara pluton.
The intrusion of the Main Donegal Granite						
The *Devlin Belt* of superimposed folds. (Related to the intrusion of the Main Donegal Granite)	DMG1:- flattening and matrix coarsening. Growth of porphyroblasts. Mullions develop.	D6:- formation of the S6 cleavage to form the dominant schistosity.	D4:- upright folds and strong NE-SW schistosity getting weaker away from the granite.	D4:- upright folds with intense NE-SW trending schistosity	D5:- upright folds and NE-SW schistosity associated with intrusion Main Donegal Granite	D6:- intense NE-SW trending foliation. Formation of upright folds and localised slides.
	DMG2:- production of new folds and cleavages. Limited development within the aureole.	Gently plunging F6 folds develop, becoming tighter towards the contact. Mullions form.				
	DMG3:- devolpment of shear belts and the cross-cleavages.	D7:- formation of conjugate cleavages	D5:-weak crenulation of S4	-	-	D7:-weak NE-SWcrenulation of S6
			D6:- conjugate crenulation cleavage	D5:- conjugate crenulation cleavage.	D6:- conjugate crenulation cleavage.	D8:- extensional conjugate crenulation cleavage
			D7:- reverse kink bands.	-	-	D9:- Reverse kink bands.

Figure 2:11:- table correlating deformation phases identified by previous workers in NW Donegal.

offsets. These were the "cross-cleavages" of Pitcher & Berger (1972). The geometry is analogous to sinistral shear bands and Type I S-C fabrics, characteristic features in the development of shear zones, (Platt & Vissers 1980). The D_8 structures are clearly visible in both the aureole and the marginal granite.

D_9:- (D_7 of Hutton 1977) produces sporadic conjugate sets of 'reverse' kink bands, (Cosgrove 1976) visible in the aureole around Lackagh Bridge and the Creeslough area, (Hutton 1977) and also occasionally in the granite, (Bunbin Hill, S of Glen).

2:5:2 Deformation within the Argyll and Southern Highlands Groups

Detailed structural interpretation, comparable to that of the Donegal Appin Group has been done in central and southern Donegal by Alsop (1987) and in mid-Ulster by Alsop & Hutton (1993), whilst the other areas, i.e. Inishowen and the Slieve League Peninisula, are awaiting re-mapping before a full regional synthesis of Caledonian deformation will be possible. The following discussion will be divided into two sections, i.e. the area to the NW of the Leannan Fault and secondly the area to the SE.

2:5:2:1 The Structures to the NW of the Leannan Fault

The rocks of this area form a narrow strip bounded to the NW by the Knockaateen Slide and to the SE by the Leannan Fault. The strata are broadly parallel to these two structures, younging towards the SE, except in the Milford area where the Upper Crana Quartzite is inverted and dips to the NW, although bounded between faults of the Leannan system, (Pitcher & Berger 1972). This structural situation led these authors to propose the rather conjectural Aghla Syncline, which faces down to the SE, associated with a penetrative, axial planar, composite S_{2-3} cleavage. Pitcher & Berger (1972) believed there was a change in facing of D_{2-3} structures across the Knockateen Slide, but this was a view not shared by later workers. On west Fanad, White & Hutton (1985) identified a similar chronology of deformation in the Argyll Group rocks to that of the Appin Group and found minor F_2 structures verge and face to the NW, throughout the area. The NW facing of these folds implies this area is on the right-way-up limb of the Aghla Anticline, casting doubt on the existence of the Errigal Syncline in this area, as Pitcher & Berger (1972) had originally suggested. In the west Fanad area F_3 verges upwards to the SE, (White & Hutton 1985). This observation was also noted by Meneilly (1982) in the Loughros Peninsula, approximately along strike, to the SW from the proposed Aghla Syncline of Pitcher & Berger (1972). Johnson *et al.* (1979) (and later Meneilly 1982) believed the Aghla Syncline to be an F_3 structure. This was based on unpublished work by Hutton who had identified NW facing structures in the Argyll rocks of the Fintown area.

Therefore the major change in the facing direction of F2 structures occurs further to the SE of the presently exposed Knockateen Slide, (Johnson *et al.* (1979), differing from what Pitcher & Berger (1972) had previously stated. Therefore, the major structure in this area is the conjectural F_3 Aghla Syncline and associated parasitic folds, with major F_2 structures apparently absent, (White & Hutton 1985). Later structures have been recorded in this area consisting of large scale open folds, (the F4b folds of Pitcher & Berger (1972), whilst in Fanad White & Hutton (1985) report the presence of gentle to open upright structures which have two dominant trends, NW-SE and NE-SW associated with rarely developed axial planar cleavage. These later folds maybe a conjugate array belonging to the same phase of deformation.

Before going on to discuss the rocks of central Donegal, the Knockateen Slide, traceable from Fanad to the northern part of the Slieve League Peninsula, will be discussed. The amount of stratigraphic displacement along the slide is regionally variable, (White & Hutton 1985). In the Fintown-Loughros area there appears to be a natural upward passage from the Creeslough Succession into the Kilmacrenan Succession, (Pitcher & Berger 1972), whilst on Fanad rocks of the Argyll Group are brought into contact with the Ards Quartzite, cutting out 1.2 km total thickness, (White & Hutton 1985). Along much of the slide there is little evidence of enhanced deformation such as mylonites, (although they are seen at Churchill and Cummirk Bridge, Hutton, *pers comm.*) which one would expect if significant displacements had occurred. White & Hutton (1985) observed dolomite breccias along parts of the slide which are interpreted as being 'carneugles'. Carneugles are original evaporite layers, that have been mobilised during burial or by tectonic processes, e.g. thrusting. When evaporites are heated, dehydration reactions will occur releasing fluids which locally generate high fluid pressures facilitating displacement without deforming the adjacent rocks to any high degree, (i.e. fluid pressure exceeds δ_3). In the French Alps such carneugles are seen along many of the major thrusts and are widely interpreted as evaporitic weak horizons which allowed tectonic translation. The age of the Knockateen Slide is uncertain, though is believed to be D_2 in accordance with the other slides in NW Donegal, i.e. thrusting to the NW (Hutton 1983). The slide separates younger rocks on top from older ones beneath, and therefore cannot be explained by simple thrusting mechanisms, (for a fuller discussion see Hutton & Alsop 1995).

2:5:2:2 The Structures to the south-east of the Leannan Fault.

The trend of the Leannan Fault System juxtaposes Dalradian rocks of differing structural trends and metamorphic grade as indicated by the high obliquity of fold axes to the SE of the fault (figure 2:7). In central Donegal the distribution of the stratigraphy is dominated by the large, sub-recumbent, NE plunging, SE closing Ballybofey Nappe, (Ballybofey Anticline of Pitcher *et al.* 1971), (Alsop & Hutton 1990), which produces a regional stratigraphic inversion over an area of 550 km^2 on the SE, translating Dalradian rocks over pre-Caledonian basement along the Lough Derg Slide, (Pitcher *et al.* 1971; Alsop 1990). Metamorphic grade increases towards the south as deeper structural levels are encountered, with mid amphibolite facies rocks occuring in the Lough Derg area. Further to the east the Sperrin Nappe, (Alsop & Hutton 1993) is carried ESE over Ordovician volcanics, by a crustal scale ductile shear zone, which intensifies towards the base of the Omagh Thrust.

2:5:2:2a Central Donegal

D_1 structures:- no unequivocal folds or lineations are assignable to D_1, due to overprinting by later phases of deformation, (Alsop 1994a *and references therein*). S_1 is preserved in the hinges of later folds where strain is lower, whilst on the limbs of these folds S_1 has been transposed parallel to bedding. The major structure associated with D_1 is the Central Donegal Slide, (Alsop & Hutton 1990; Alsop 1992c). The rocks to either side of this slide consistently young away from this structure leading Alsop & Hutton (1990) to construct tectono-stratigraphic successions for the rocks of central Donegal. In the footwall of the slide (prior to later folding) occurs the inverted sequence of rocks belonging to the Finn Succession, whilst in the hanging-wall there are right-way-up rocks of the Ballybofey Succession, (Alsop & Hutton 1990). The slide initiated at greenschist facies, (though reactivated by later deformation at amphibolite facies) and is related to NW directed thrusting, although D_1 shear sense is no longer visible due to later overprinting (Alsop 1992c). The reversals in younging on either side of the slide are interpreted as the formation of a north-west facing, isoclinal, recumbent D_1 fold with the hinge cut-out by the propagating thrust.

D_2:- in central Donegal S_2 is a prominent fabric consistently dipping towards the NE, although in many areas it has been extensively modified by D_3. In the hinges of F_3 folds the crenulation of S_1 by S_2 is clearly visible, (Alsop 1994*a*). On the upper limb of the Ballybofey nappe minor F_2 folds are visible, being tight to isoclinal, semi-recumbent folds, but are generally rare, (or difficult to identify, becoming intra-folial due to later high strains), Alsop (1994*a*). In the hinge of the D_3 Ballybofey Nappe, S_2 is clearly folded by this later structure, whilst on the inverted limb of the nappe increasing D_3 strain results in S_2 becoming parallel to S_3 forming a composite N, NE dipping S_{2-3} cleavage, whilst F_2 folds become rare to absent, (Alsop 1994*a*).

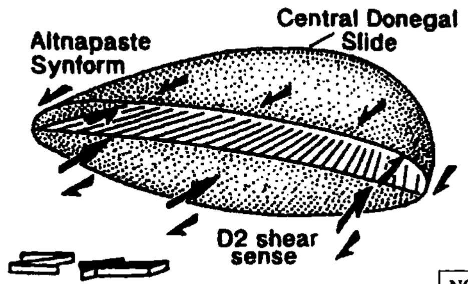
In the hinge of the Ballybofey Nappe minor F_2 sheath folds are visible and locally these structures change vergence across F_2 axial traces, defining the presence of kilometre scale sheath folds of the same age.

Lithostratigraphical evidence also indicates the presence of these folds, e.g. the sub-circular outcrop of the Gaugin Quartzite, (although truncated by the later Belshade Fault) within the hinge of the Ballybofey Nappe. This outcrop relationship is related to the presence of the D_2 Altnapaste Synform, (Alsop, 1994*a*), a recumbent isoclinal structure closing towards the ESE. L_2 stretching lineations trend ESE-WNW, with the overshear direction indicated by the closure of the sheath fold, i.e. towards the ESE. West of the F_2 Altnapaste Synform, Alsop (1994*a*) recognised another F_2 fold due to:- (a) the outcrop geometry of the Reelan Formation and (b) a reversal in younging across the axial surface. This is the Croveenananta anticline of Pitcher *et al.* (1971), which they believed to be an F_1 structure. The development of these structures in this area may have been due to the reactivation of the pre-existing Central Donegal Slide, an anisotropy which localised strain during D_2 deformation and possibly caused attenuation of F_2 folds to produce sheath geometries, (see figure 2:12a). D_2 shear sense along the Central Donegal Slide reveals top to the SE movement. On the inverted limb of the later Ballybofey Nappe the F_2 folds become strongly overprinted by D_3 events, notably by movements along the D_3 Ballybofey Slide, (Alsop 1992d). The coaxial and coplanar nature of D_2 and D_3 creates difficulty in identifying D_3 modification of F_2 structures, i.e. D_3 may have enhanced the sheath-like nature of these folds.

D_3:- This is responsible for the major recumbent, SE facing Ballybofey Nappe and associated tectonic slides forming Type II interference folding with D_2 folds, (figure 2:12b). D_3 is associated with oblique dextral translation of the nappe, (figure 2:12c) towards the ESE, with the strain gradient increasing towards the base of the nappe, (Alsop 1994b).

On the upper limb S_3 is a gently, NE dipping, axial planar schistosity which possesses an axial planar relationship to tight to reclined F_3 folds. These folds usually verge towards the SE and towards the hinge of the Ballybofey Nappe, (Alsop, 1994b). The orientation of F_3 fold axes changes down through the nappe, in response to an increasing D_3 strain gradient as lower structural levels are encountered. Moving towards the hinge of the nappe the D_3 strain increases, in response to localised reactivation of the Central Donegal Slide and the initiation of the Ballybofey Slide, (Alsop, 1992d).

On the inverted limb D_3 intensifies, transposing S_2 to form a composite cleavage, with L_{2-3} also, forming a composite lineation. In the foot-wall to the Ballybofey Slide there is D_3 synformal complex, (the Silver Hill Synformal Complex) which folds the Ballybofey Slide. This relationship proves the progressive and protracted nature of D_3 deformation with ductile thrusts being folded by structures assigned to the same age. The entire overturned limb of the Ballybofey Nappe is a 10 km thick shear zone within which D3 strain increased downwards, with the development of coeval ductile thrusts, i.e. the Ballybofey Slide and the Finmore Slide, eventually culminating with the Lough Derg Slide, which forms the basal decollement to the nappe: thrusting the Dalradian over pre-Caledonian basement in an oblique dextral manner towards the ESE, Alsop (1994b). Within the Lough Derg inlier the amphibolite facies D3 thrusting event has created an intense S3 foliation and foot-wall synform; the Lough Derg Synform, (Alsop 1991).



(A) D2 sheath folds due to reactivation of Central Donegal Slide (Alsop 1994*a*)

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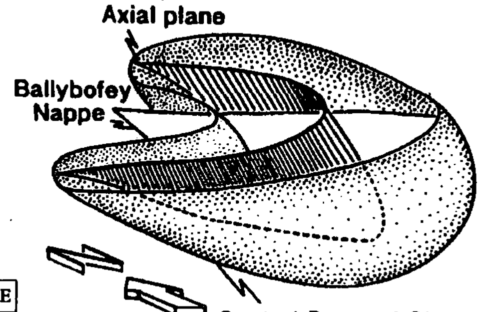
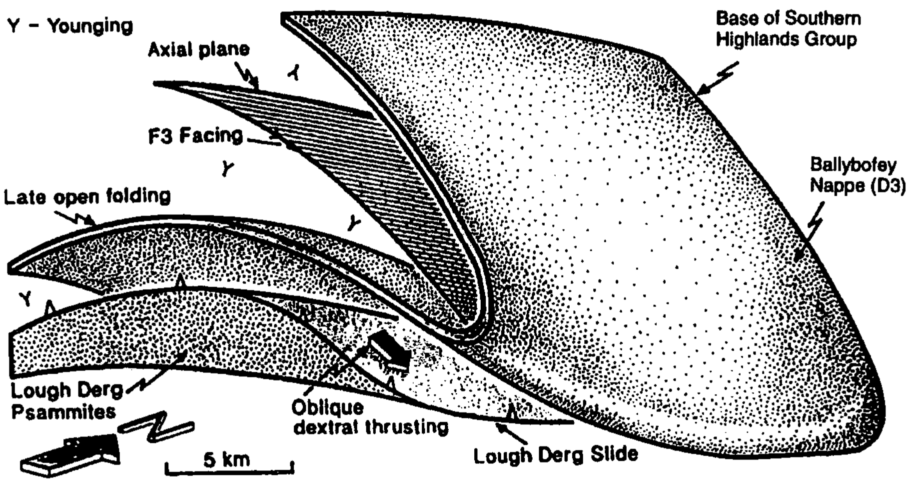


(C) Refolding of D2 structures by the Ballbofey Nappe (Alsop 1994*a*)



(C) The Ballybofey Nappe (Alsop 1994b)

Figure 2:12:- Schematic diagrams of the main structures within central Donegal (after Alsop 1994a,b)

Subsequent to nappe emplacement and associated crustal thickening, there is an episode of ductile extension, visible in the hanging walls of both the Ballybofey Slide and Lough Derg Slide, (Alsop 1991; 1992d). Extension is orthogonal, (i.e.

down dip) on S_3 surfaces and is defined by a well developed stretching lineation, plunging in a northwards direction. S-C fabrics also testify to down dip extension towards the north. The synchroneity of crustal thickening and subsequent extension is indicated by kyanite porphyroblasts which grew in response to the former event, but also defines the stretching lineation associated with the extensional event, (Alsop, 1991).

The Ballbofey Nappe was later folded by open, upright F_4 folds in response to later E-W compression with the formation of the Ballard Antiform and further to the south, the Lough Derg Antiform, both of which post-date the ductile extensional event.

2:5:2:2b Eastern Donegal and mid-Ulster

Early workers believed the rocks of mid-Ulster to be on an inverted limb of a major, gently inclined, recumbent, SE facing F_1 fold, later termed the Claudy Anticline by Pitcher & Berger (1972) who believed this structure to be the equivalent of the Tay Nappe in SW Scotland. Remapping of much of this area by Hutton, (*see* Hutton & Alsop 1993b) showed that much of the Dalradian stratigraphy in the northern Sperrin Range was the correct way-up, though a major inversion of stratigraphy was encountered further to the south in the Sperrin range. This allowed a relocation of the axial trace of the major SE facing nappe from Claudy to the Sperrins, hence the Sperrin Nappe, (Alsop & Hutton 1993b). The inverted rocks on the lower limb of the Sperrin Nappe extend southwards towards the Omagh Thrust, where the nappe is carried over Ordovician volcanics in a east-south-easterly direction, (figure 2:13).

In this area S_1 is everywhere parallel to bedding with no F_1 folds identifiable. D_2 events form prominent structures within the area, with S_2 being an intense transposing crenulation cleavage (Alsop & Hutton 1993b) On the upper limb of the Sperrin Nappe minor folds are tight to isoclinal, (with a moderately NNW dipping axial planar cleavage) indicating that before D_3 facing and vergence of D_2 structures was to the SE. The reversal of younging and change in vergence of F_2 structures across the axial trace identifies the Sperrin Nappe as D_2. On the inverted limb F_2 structures becoming progressively rarer due to later overprinting D_3 deformation. D_2 in this area is associated with mid-upper greenschist facies metamorphism, (Alsop & Hutton 1993b).

Within the Sperrin Nappe there is a D_3 strain gradient which increases in a SE direction as deeper structural levels are encountered. On the upper limb of the Sperrin Nappe, minor tight F_3 folds range up to a kilometre in wavelength and are recumbent

F_3 folds including the Butterlope Fold pair which locally inverts the stratigraphy, and verge to the west, (Alsop & Hutton 1993*b*).

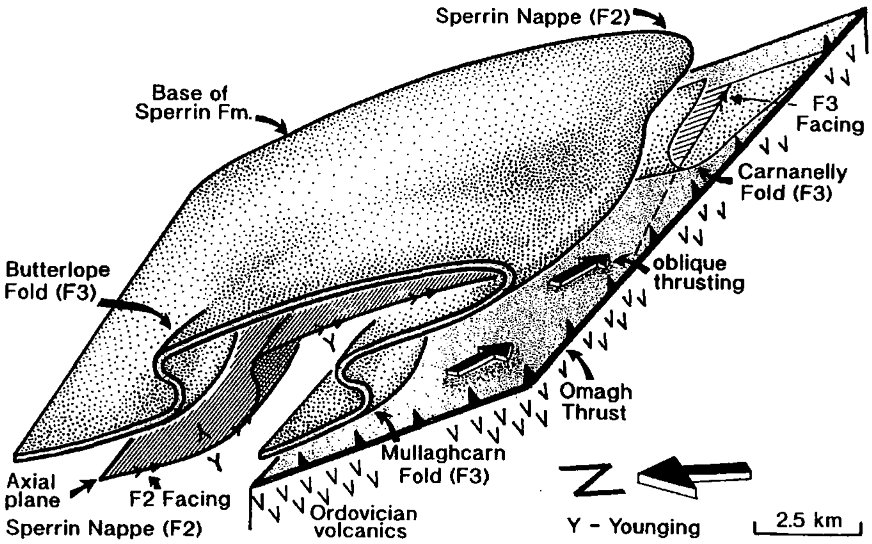


Figure 2:13:- The structure of the Sperrin Nappe, mid-Ulster (*after* Alsop & Hutton 1993b)

On the inverted limb of the nappe S_3 is sub-parallel to S_2 becoming indistinguishable in the field, forming a composite S_{2-3} foliation. On this limb there are large F_3 folds including the Mullaghcarn fold pair which locally restores the beds to right-way-up on part of the limb, (Alsop & Hutton 1993b). Also present is the westerly verging Carnanelly Fold which reduces in amplitude and significance upwards through the nappe, (see figure 2:13).

As already mentioned the D_3 strain increases towards the base of the Sperrin Nappe, intensifying into the Omagh Thrust, with the development of extensional crenulations displaying top to the ESE ductile thrusting, (Alsop & Hutton 1993a). The lower greenschist facies Ordovician volcanics preserve a fabric correlated with D_3 in the Dalradian of Alsop & Hutton (1993*b*) extending 2-3 km south from the thrust contact. In summary D_3 records the development of a 10 km thick shear zone which was responsible for translating the D_2 Sperrin nappe over Ordovician rocks to the south. Northwards the D_3 strain diminishes to preserve D_2 strains in the upper limb of the nappe, which originally had a N-S stretching lineation associated with southerly directed thrusting.

Alsop & Hutton (1993a) identified extensional structures along the Omagh Thrust and in the hanging wall attributable to gravitational instability in response to the crustal thickening events of D_3. These down dip extensional structures are very similar to those at the base of the Ballybofey Nappe.

2:6 Summary of the pre-granite geology of Donegal

Before going on to discuss the granites which comprise the Donegal Batholith the geology of Donegal will be summarised

The failure in this area to identify any D_1 structures in NW Donegal, comparable with the NW facing F_1 Islay Anticline of Scotland, led Hutton (1983) to suggest two possibilities, although he favoured the latter option:-

- the Appin Group in this area is on the upper limb of a major, recumbent, NW facing nappe of which the hinge and inverted limb are below sea-level.
- the hinge of the nappe has been cut-out by subsequent tectonic sliding.

On Inishowen major D_1 structures are preserved with the preservation of a large upright syncline which is believed to be the Irish equivalent of the Loch Awe Synform visible in SW Scotland (Hutton *pers. comm.*).

In central Donegal, Alsop (1992c), interpreted the Central Donegal Slide as a D_1 structure with the rocks on either side consistently younging away from the slide, implying that sliding must be related to a major recumbent fold responsible for inverting the footwall rocks. The Central Donegal Slide is believed to have been a NW directed thrust in which the hinge of the associated recumbent fold was excised. Further north Alsop, (1991) postulated that part of the Knockateen Slide is of similar age. Although White & Hutton (1985) believed there was no change in facing of F_2 structures across the Knockateen Slide, Alsop (1991c) reported a facing change in the area around Glenties, which would suggest D_1 movements on this structure. The presence of this structure near the strike-swing in Donegal also suggests that earlier deformation nucleated in this area, as suggested by Hutton & Alsop (1996).

D_2 deformation is related to the emplacement of foreland propagating nappes carried north-westwards by a series of ductile thrust-slides. The correlation of D_2 between NW Donegal and central and southern Donegal is difficult. In central Donegal F_2 sheath folds are developed in the hinge area of the later Ballybofey Nappe and are related to translation to the SE. Reactivation of earlier structures also occurs (e.g. the Central Donegal Slide, (Alsop 1994a)). In mid-Ulster the Sperrin Nappe is associated with southerly directed translation, (before D_3 deformation).

In NW Donegal structures produced by D_3 show an important reversal in the facing direction and vergence towards the SE, when compared with D_2 (Hutton 1977). In the northern area of the Appin Group the intensity of D_3 is limited, becoming more dominant towards the south in higher stratigraphical levels with the development of recumbent folds. Hutton (1983) suggested the D_3 event was the result of gravity controlled back folding in response to the D_1 and D_2 thrusting events.

In central and southern Donegal, D_3 deformation forms the major structures, e.g. the Ballybofey Nappe. On the inverted limbs of the Ballybofey Nappe and the Sperrin Nappe there are crustal scale D_3 shear zones in which strains increase downwards towards ductile thrust-slides (the Lough Derg Slide and the Omagh Thrusts) which have oblique dextral movements. Alsop (1994b) showed that with the removal of late stage folds (e.g. the D_4 Ballard Antiform and Lough Derg Antiform) the Ballybofey Nappe appeared to have originally faced slightly downwards to the SE, allowing an origin for this structure by gravity controlled collapse.

Immediately following the crustal thickening event there was extensional collapse down to the NW within the inverted limbs of the both the Ballybofey Nappe and the Sperrin Nappe. Alsop & Hutton (1993*a and references therein*) related this collapse to dynamic wedge models for orogenic belts where fluids released during prograde metamorphism associated with crustal thickening induce collapse due to high fluid pressure along structural anisotropies. The D_3 event in central Donegal has been dated from deformed pegmatites within the Lough Derg Inlier, which give Rb-Sr muscovite ages of 460 Ma This may give a cooling age associated with amphibolite facies which is broadly the same age as the D_3 event in Ulster where faunal evidence and synchronous volcanism with deformation suggest an Arenig-Llanvirn age, (Alsop & Hutton 1993*a*, *and references therein*). Whether or not the D_3 of NW Donegal, (Hutton 1977) is the same as the D_3 of central Donegal is still an area for future research. Although the Leannan Fault System is of post-granite age, it is convenient to mention it here as one of the major structures in Donegal.

The Leannan Fault juxtaposes Dalradian rocks of differing structural trends and metamorphic grade, (Alsop 1992a). Movement along similar, although smaller faults are further constrained by their truncation of some of the Devonian granites. The Barnesmore granite, 397 ±7 Ma, (O'Connor *et al.* 1987) is cut by the Belshade Fault, (Walker & Leedal 1954) which it sinistrally displaces, together with a component of down to the SE dip-slip. The continuation of the fault into the Carboniferous Sandstones in southern Donegal only shows the dip slip component, putting an upper limit of Visean, (approx. 352 Ma) (Alsop 1992a *and references therein*) to the sinistral component to these faults.

Alsop (1992a) traced the upright Foyle Synform across the Leannan Fault to give a sinistral displacement of 34 km. Comparison of lateral facies and metamorphic grade after restoration shows good correlation with this value and also the Ballybofey Nappe aligns with the Glencolombkille Antiform, implying they were originally the same structure (Alsop 1992a *and references therein*). Previous estimates of displacement have greatly varied, e.g. 40 km, (Pitcher *et al.* 1964; Pitcher & Berger 1972), and 160 km, (Phillips & Holland 1981). The fault was believed to be the extension of the Great Glen Fault into this area, (Pitcher *et al.* 1964), though with greatly diminished displacement compared to values observed in Scotland. The increasing geophysical coverage of the area around the north Irish Sea identified the Great Glen Fault to the north of the Irish mainland, implying the Leannan Fault is a major splay of the Great Glen Fault System, (Pitcher & Berger 1972).

Chapter 3

The Donegal Batholith: an introduction to the Main Donegal Granite

3:1 The Donegal Batholith: an introduction.

The Donegal Batholith within which eight plutons have been identified, has been an area of extensive study, notably over the past 50 years (Pitcher & Berger 1972). The age of the Batholith is reported to be in the range 418-397 Ma (Rb-Sr), (O'Connor *et al.* 1982), (see figure 3:2 for individual ages) with the plutons showing a wide variety of emplacement styles at a similar crustal level (Pitcher & Berger 1972). The depth of emplacement of the Donegal granites, with the exception of the Barnesmore pluton, was approximately 2.5 Kb (using garnet-plagioclase-Al_2SiO_5-quartz geobarometry for the Ardara pluton, (Kerrick 1987)) equating to a depth of approximately 10 km. Hutton (1982) proposed a unified tectonic model for the emplacement of the Donegal Batholith in which a sinistral shear zone, along part of which the Main Donegal Granite was emplaced, structurally controlled the siting of other plutons. Within shear zones localised zones of extension and compression can be generated in fault tips, restraining bends and releasing bends (Sanderson & Marchini 1984), which may control the manner of emplacement, i.e. permissive or forceful. The Thorr and Fanad plutons were permissively emplaced by stoping in a dilational zone between two shear zones, the Main Donegal Granite Shear Zone, (MDGSZ) in the south and an inferred shear zone to the north, possibly a member of the Great Glen Fault System (Hutton 1982). The Ardara and possibly Toories granites were emplaced in a compressional domain, (such as a shear zone tip) by ballooning mechanisms. Hutton (1982) proposed that the Main Donegal Granite "wedged" itself into the shear zone in sheet-like masses, causing the shear zone to split axially. Zones of extension due to bending of one side of the shear zone allowed the permissive emplacement of the Rosses Pluton by cauldron subsidence and Trawenagh Bay by stoping mechanisms. Watson (1984), suggested the presence of a NW-SE lineament, centred on the Ardara granite.

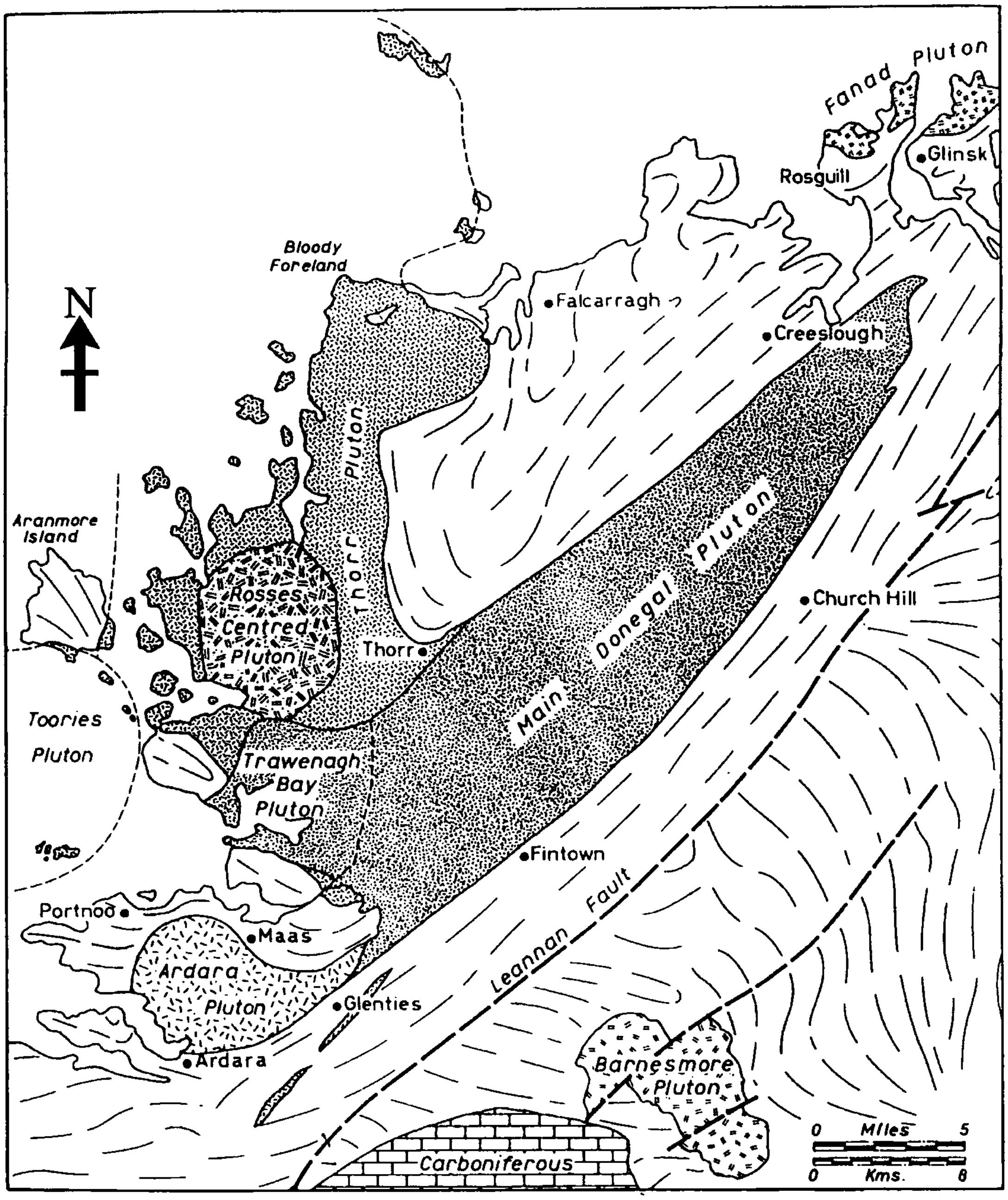


Figure 3:1:- The Donegal Batholith (*after* Pitcher & Berger 1972)

This is analogous to similar trending lineaments in Scotland, all of which controlled granite emplacement, e.g. the Loch Shin Line, the Cruachan Lineament and Strath Ossian Lineament, (Jacques & Reavy 1994). This model was revised by Hutton & Alsop (1996) who proposed the siting of five of the eight plutons along a reactivated NNE-SSW striking pre-Caledonian lineament, which had previously controlled the deposition of the Dalradian sediments and the Caledonian strike swing. At the intersection of the lineament and the MDGSZ is the Ardara pluton and a host of appinitic complexes.

The order of emplacement of the plutons, based on cross-cutting relationships is Thorr, possibly Fanad, Ardara and Toories, Rosses, Main Donegal Granite with the petrographically similar but weakly deformed Trawenagh Bay Granite, (Pitcher & Berger 1972). The age relationship of the isolated Barnesmore Pluton is uncertain but the Rb-Sr isochron data unequivocally relates in time to the other members of the batholith, (O'Connor *et al.* 1987) (figure 3:2).

Pluton	Age (Myr (2σ)) Rb-Sr (whole rock)	Initial $^{87}Sr/^{86}Sr$	Initial εNd (Dempsey *et al*, 1990)	References
Thorr	418 ±26	0.7057	-4.8	
		0.7051	-5.1	O'Connor *et al*, (1982)
Fanad	402 ±10	0.7050	-	O'Connor *et al*, (1987)
Ardara	405 ±5	0.7062	-1.2	Halliday *et al*, (1980)
Toories	-	-	-	-
Rosses (G1)	404 ±3	0.7066	-8.0	Halliday *et al*, (1980)
(G3)		0.7062	-4.9	
Main Donegal	388 ±3	0.7062	-8.2	Halliday *et al*, (1980)
	407 ±23	0.7058	-8.3	O'Connor *et al*, (1982)
	407 ±4 *			O'Connor *et al*, (1984)
Trawenagh Bay	405 ±3	0.7068	-7.5	Halliday *et al*, (1980)
Barnesmore, G2 & G3	397 ±7	0.7065	-4.7	O'Connor *et al*, (1987)
		0.7057	-4.3	

* *207Pb/206Pb age dating from uraninite*

Figure 3:2:- Rb/Sr isochron data and initial εNd values for the Donegal Granites.

The source to some of the Donegal Granites can be inferred to some degree by the ε_{Nd} data. The more highly negative values within some of the plutons suggest the involvement of very old LREE enriched crust e.g. basement. The Rosses (G 1), Trawenagh Bay Granite and Main Donegal Granites all have highly negative ε_{Nd} values (figure 3:2). The less negative values implies that juvenile crust or mantle material has contributed to the petrogenesis of these granites e.g. Thorr and Ardara. The highly mafic nature of these plutons suggests some mantle contribution. The ε_{Nd} for the Barnesmore Pluton is similar to the Thorr pluton. This former pluton consists of relatively highly evolved granites, inconsistent with mantle or basic component, unless they are highly fractionated residues from the same source. Dempsey *et al.* (1990) thought that the Barnesmore Pluton may have different basement beneath it in comparison to the other members of the Donegal Batholith. Furthermore the pluton in question lies to the south of the Leannan Fault, across which 35 km of sinistral displacement has been recorded (Alsop 1992a). Displacements of this scale could juxtapose younger basement into this area, which would account for the intermediate ε_{Nd} values. No ε_{Nd} data exist for the Fanad pluton although the very high strontium content may imply some mantle origin. At mantle depths, plagioclase is unstable making Sr a highly compatible element (Dempsey *et al.* 1990). Halliday *et al.* (1985) suggest 50% mantle derived material and 50% lower crustal material were involved in the generation of the Fanad monzodiorites.

3:2 Previous Work on the "Donegal Granite"

This section describes some of the observations made by early workers in Donegal before the detailed mapping programme of the Imperial College Research Team, (post 1948), a project initiated by Professor H.H. Read to study a well exposed granite batholith and its adjacent host rocks.

The following résumé is taken from a thorough account given by Berger, (Unpublished Ph.D Thesis, University of Liverpool, 1967) and Pitcher & Berger (1972). The earliest geological map of Donegal was produced by Griffith, (1839), who observed a "v-shaped" mass of granite, (the later differentiated plutons: Thorr, Rosses, Trawenagh Bay, Ardara and the Main Donegal granites), within a NE-SW trending belt of meta-sediments. The production of this map provided a stimulus for further research by later workers. The first report on the Donegal Granite was by Kelly (1853), who suggested an igneous origin for the foliated granite from the distribution of the major quartzites overlain by "*mica-slate*". Kelly interpreted the granite as occupying the core of an antiform with the overlying quartzite, either having been eroded due to the uprising granite **or** it was incorporated into the granite. Over the next forty years the origin of the "Donegal Granite" was to be the subject of great debate, i.e. was it produced by an intruding magmatic liquid or by in situ transformation of the country-rocks. In hindsight this early debate was greatly complicated by the failure of these workers to recognise individual members of the Donegal Granite, (listed above) which, as it was discovered a hundred years later, possessed strongly contrasting emplacement phenomena. In 1861, Scott and Haughton were undertaking active research on the granites, and in 1862, Scott reported "*the typical Donegal Granite presents no appearance of being a purely igneous rock, the evidence in fact pointing to a metamorphic origin for it*" and was "*formed from materials existing on the spot without actual fusion*". The presence of meta-sedimentary inclusions and the gneissose character of the granite led Scott (1862), Blake (1862) and Haughton (1862) to suggest a metamorphic origin. Blake commented on the lack of mineralisation within the Donegal Granite, a feature usually characteristic of granites of an "irruptive" origin. The following year, Haughton & Jukes (1863) studied the gneissose character of the Glen-Gweebarra granite and their observations led Scott to suggest a similarity with the Laurentian Gneisses of Canada. In 1863 Scott summarised the conclusions of Haughton and himself and stated how Haughton, despite having performed chemical analyses on the Donegal Granites and other Caledonian plutons had given little consideration to the origin of the granites. The exception to this was a statement about the central portion of the foliated granite, (Main Donegal) "*..perhaps of igneous origin, originally deriving its cleavage planes and the gneissose character from the pressure exerted on it from the north and south, to a nearly vertical position.*" This now might be interpreted as uprising granite pushing against confining pressures in the country rocks resulting in the development of a foliation within the granite.

The next major contribution to Donegal geology was made by Hull and Kinahan in the 1880's. Hull (1881; 1882), supported Scott and Haughton's conclusions that the gneissose granite was a "metamorphosed stratified formation", thus re-igniting Jukes's belief that the granites were Laurentain in age and the Dalradian meta-sediments unconformably overlay the granite gneisses. Hull also classified the xenolith-free granites near Dunglow, (the Rosses and Trawenagh Bay) as deeper granites where "*fusion had superseded metamorphic action*". The xenolith rich granites (Thorr and Main Donegal) were regarded as lateral equivalents where heat had not been sufficient to totally fuse the material, leaving foliated granite inter-stratified with schistose meta-sediments.

In 1885, work on the Donegal Granites was greatly influenced by Callaway. At many localities in Donegal he observed granitic material truncating the strong foliation within the schists, implying an intrusive origin. In the Barnes Gap, Callaway wrote "*where I first touched the granite, it was perfectly clear that I was on the margin of an intrusive mass*". It should be recorded that the present author first "touched" the Main Donegal Granite at this exact locality, where the sheeted relationships of granite into country-rock, as reported by Callaway, are superbly exposed. Callaway strongly rejected the metamorphic origin for the granite put forward by Scott, Haughton and Blake, and proposed the following evidence for an igneous origin:-

i) The gneissose character and inclusions of schist were still compatible with an igneous origin.

ii) The occurrence of granite "beds" parallel to the schists was due to the intrusion of granite along the bedding planes, as indicated by minor irregularities along the contacts.

iii) There is no transition from schist to granite, a phenomena one would expect to observe for a granitisation type process. Callaway observed that the contacts were mainly sharp, although slightly "welded".
iv) The uniformity of the granite "*could not have been produced by the fusion of a varied series of strata*".

The above observations are a powerful demonstration of the importance of detailed observational fieldwork when presenting a legitimate argument. As Read said in the discussion of Pitcher & Read (1959) "*the facts would remain the same however the interpretation may change*".

The work of Callaway led Hull (1886) to study the evidence a little more carefully and resulted in the latter geologist suggesting that the granites of Donegal "*might belong to at least two periods- the intrusive being distinct in both age and structure from the metamorphic granites and gneisses*". Slowly the "transformers" of Hull and Kinahan were being converted to "magmatists", for at least some areas of the "Donegal Granite".

In 1891 the Geological Survey Memoir on North-western Donegal (and maps on a scale of 1 inch: 1 mile) had been published by Hull, Kilroe & Mitchell. Kinahan had refused to be acknowledged as an author, because he thought Hull had changed many of his descriptions so they would fit observations of Hull. The argument stemmed from the absence of adequate way-up criterion, which prevented the construction of a stratigraphic succession for the meta-sediments.

The Director General of the Irish Geological Survey at this time, Sir Archibald Geikie commented on the similarity between the rocks of Donegal with those of central and SW Scotland and termed this diverse rock sequence the "Dalradian", (named after the ancient kingdom of *Dalradia* which extended over both areas). Hull, in the memoir, changed his earlier beliefs about the age of the granite-gneisses, saying that he no longer thought they were Laurentian and that the Dalradian lay above, instead suggesting the granites were a later "*irruptive mass*" and "there were at least four or five periods of granite intrusion". He stated that his "unconformity" was now seen to be a "*transgression of the granite across the rocks*" with the foliated granite of Glenveagh traceable "*with uninterrupted passage*" into the unfoliated granite of Dunglow and Trawenagh Bay, where the granite could clearly be seen truncating quartzite beds. Hull believed the foliated structure of the central region was created by lateral thrusting and shearing movements, "*formed after the main foliation of the region had been completed*". This belief was confirmed by Pitcher & Berger (1972), Chenevix-Trench (1975) and Hutton (1982) almost a hundred years later. Kilroe in the same memoir concluded that the "*intrusion of the granite marks an important* *epoch in the geological history of the Donegal area, having occurred, it would seem between earlier and later periods of metamorphism*".

In the early part of the 20th Century, Grenville Cole compiled a series of papers about the rocks of NW Donegal. Cole (1902) agreed with the views of Callaway about the intrusive nature of the Donegal Granites. Cole stated that the granite foliation was due to the effects of flow during emplacement and that this was "*emphasised to some extent by subsequent pressure*". He worked in western Donegal, in the area covered by the old estate of Boylagh and suggested that the granite of the Fintown-Galwollie Hill area was intruded into the core of a large anticline. The "serrated" granite contact of Derkbeg-Straboy area, where country-rock inclusions are abundant, was interpreted by Cole as roof-pendants incorporated into the rising granite body. In 1906 and 1913 he extended these conclusions saying the inclusion rich granite-gneiss was the result of a rising granite dome invading "previously foliated sedimentary and igneous material". He further wrote "whole masses of granite may thus invade and take the place of a pre-existing schistose series without destroying all traces of the of the original structure of the district". Cole believed this was the extreme case of "lit-par-lit" injection.

Over the next thirty years, work in Donegal was limited to a number of minor contributions to the literature, e.g. Andrews (1928) who had studied the margins of the Donegal Granite and supported a *lit-par-lit* injection mechanism for these areas and further stated the gneissossity was not created by post-consolidational shear, but due to "flow" of the crystallising magma.

From the above résumé one can appreciate how controversial was the question of the origin of the "Donegal Granite" between the 1840 and 1945, with much of the debate revolving around the question of whether the granites were of metamorphic or magmatic in origin. To summarise Kelly (1853) favoured the igneous origin for the granites, whilst Scott, Haughton & Blake preferred the metamorphic view during the 1860's. In the early 1880's Hull and Kinahan continued the metamorphic hypothesis, though they were later converted to magmatists by the work of Callaway who strongly favoured the igneous origin. Work in the early years of this century by Cole and Andrews continued to find evidence to support the igneous origin for the Donegal Granite, (or at least parts of it).

During the 1940's and 1950's the arguments concerning the origin of granite were to re-emerge again. The debate centred on the fact that some granites, at the same level of exposure, showed evidence of both granitisation-migmatisation and actual intrusion. To clarify this debate H. H. Read thought that the well exposed granites of Donegal would be an ideal area to test his depth controlled granite series model, (of Read 1948). The Granite Series was "*an attempt to relate plutonic* *phenomena at the various levels of exposure and to give unity to the processes of granitisation, migmatisation and metamorphism at depth and successively at higher positions at later times*". The first students of the Imperial research team started in 1948 and began to re-map individual areas of Donegal on a scale of 6 inches: 1 mile. This mapping continued for the next 20 years, with the supervision of the work being later taken over by W.S. Pitcher, one of the initial members of the research team. Relevant members of the research team include, (also see references): A. R. Gindy; W. S. Pitcher; G. J. H. McCall; I. C. Pande; E. H. T. Whitten; S. V. P. Iyengar; C. F. Tozer; E. L. P. Mercy; R. L. Cheeseman; M. K. Akaad; D. C. Knill & J. L. Knill, T. C. R. Pulvertaft, M. J. Rickard (1962), F. W. Cambray: A. R. Berger and R. S. R. Wood. The work of these scientists provided a large database which allowed Pitcher & Berger (1972) to compile and construct a synthesis for the emplacement and form of the Donegal Granites. More recent work has been carried by:- J. R. Chenevix-Trench; D. H. W. Hutton; A. N. Meneilly; G. I. Alsop, M. A. McErlean, S. J. Jolley and S. M. Molyneux. Geochronological work on the Donegal Granites has been performed by Halliday *et al.* (1980); O'Connor *et al.* (1982; 1984; 1987) and Dempsey *et al.* (1990).

3:3 The Donegal Granites

The next section describes some of the main features of the granites which comprise the batholith while the latter part of this chapter is devoted to a more thorough discussion of the Main Donegal Granite. The granites will be addressed in approximate order of age.

3:3:1 The Thorr Pluton

The Thorr Pluton is believed to be the oldest of the Donegal Granites; a relationship based on cross-cutting relationships with the other plutons, notably Rosses, Main Donegal and Trawenagh Bay (Pitcher & Berger 1972). Dating of the pluton by Rb-Sr (whole-rock isochron) methods also suggests it is the oldest with an age of 418 ±26 Ma, (O'Connor *et al.* 1982). The pluton occupies an area of approximately 800 km^2, of which a large part is submarine to the N, NW and W of the mainland exposure (Evans & Whittington 1976) (see figure 3:3).

To the south of the Main Donegal Granite there is an exposure of intensely deformed tonalite-diorite known as the "Carbane Gneiss" which is petrographically similar to the granitoids which make up the Thorr pluton, (Cole 1902; Pitcher & Berger 1972). The age of the Carbane Gneiss is 415 ±3 Ma, (Rb-Sr whole-rock isochron; O'Connor *et al*, 1982), making it one of the older intrusions of the Donegal Granite Complex.

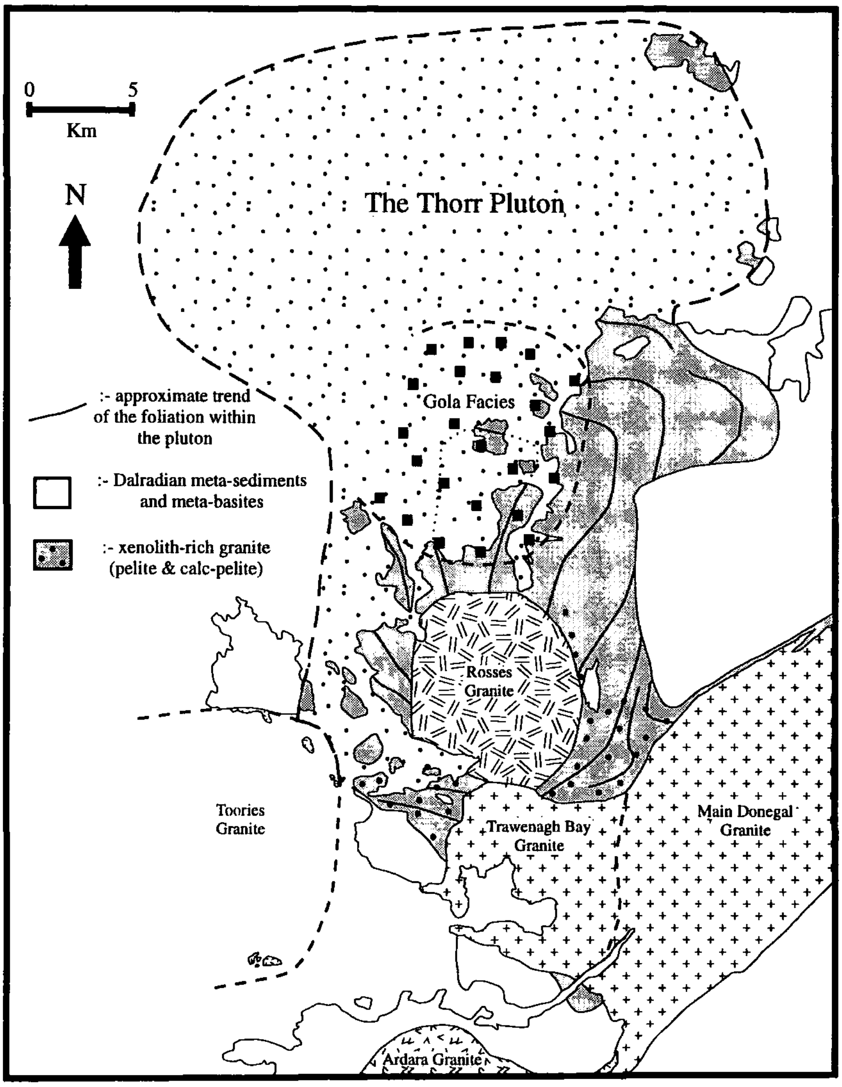


Figure 3:3:- The full extent of the Thorr pluton (NE prolongation omitted), (*taken from* McErlean 1993).

The Thorr pluton is crudely zoned from the core to the margin. The central unit consists of felsic, alkali feldspar rich granite in the central area is known as the "Gola Facies", (Whitten 1957b; Pitcher & Berger 1972). Towards the margins, especially in the south, the granite becomes more granodioritic, tonalitic and dioritic. Xenoliths of the Thorr pluton are abundant within the Main Donegal Granite, implying the existence of a NE extending "prolongation" of the Thorr pluton in this area prior to the emplacement of the Main Donegal Granite. The mineral alignment within the main body of the pluton is generally weak and is broadly parallel to the contacts, although in localised areas the trend of this fabric is discordant. McErlean (1993) identified more subtle sub-fabrics based on different alignments of different mineral phases which implies syn-magmatic sinistral shear (see section 1:4:1, figure 1.6b).

Xenoliths of country rock, ranging in size from a few centimetres to 1600 metres, are very abundant in the south and south-west of the Thorr outcrop but become progressively rare and eventually absent in the northern part of the mainland exposure (Pitcher & Berger 1972; McErlean 1993). The distribution of the xenoliths forms a "ghost stratigraphy" across the pluton with the xenoliths belonging to the formations of the Appin Group above the Ards Quartzite (Pitcher & Berger 1972). The same authors believed the rafts depicted a northward plunging anticline, the Thorr Anticline, whose axial trace was parallel to that of the folds in the envelope. Furthermore, the strike-swing can be traced through the rafts, implying this feature was of pre-granite age.

The granite can be divided into a number of facies, i.e. the normal facies, Gola facies and contact facies:-

The normal facies comprises the marginal areas of the pluton, occupying most of the present day outcrop, ranging in composition from granite *(sensu stricto)* to quartz-diorite. The petrography includes hornblende, plagioclase, (oligoclase, although anorthite content increases with colour index (Pitcher & Berger 1972), biotite, K-feldspar and quartz plus a host of accessories.

The Gola Facies has a similar grain size to the normal facies, but contains no hornblende, has fewer accessory minerals and the plagioclase is more albitic. There is no directly observable contact between the two facies

Marginal and Contact Facies:- seen at the periphery of the pluton in contact with the envelope rocks and around country rock xenoliths within the pluton. Although highly variable depending on what the composition of the country rock is, this type of granite is seen in the Maghery and Lettermacaward areas in the SW part of the pluton. Pitcher & Berger, (1972) stressed the difference between the appearance of xenoliths in the contact facies from those in the marginal normal facies. In the normal facies pelitic xenoliths are usually recognisable implying mechanical incorporation into the magma, whilst in the marginal facies the pelitic material is preserved only as mafic aggregates implying some chemical reaction, i.e. localised assimilation. Often these marginal granites are variable in composition and grain size and contain biotite, muscovite, garnet and sometimes tourmaline, plus higher amounts of quartz, suggesting per-aluminous type affinities. Around countryrock rafts, (pelite, semi-pelite, quartzite and metadolerite) the granites show localised evidence of assimilation, though the width of these "contact" granites depends upon the composition of the host xenolith. These “contact” granites are best developed around rafts of pelitic material where they can be up to 5-10 metres in thickness, being less developed around semi-pelites and meta-dolerites and almost absence around quartzites. The width of the contact facies also demonstrates that wholesale contamination has not occurred, as geochemically proven by Oglethorpe (1987), (*in* McErlean 1993).

The envelope to the Thorr pluton consists mainly of quartzite in the north and western areas, whilst later granites flank the pluton in the south. Limited strips of pelite forms the aureole in the SW (Lettermacaward area) and in the east (Thorr-Meencorwick area). The thermal metamorphic event related to the intrusion of the Thorr pluton has been described as a "static" event, (Pitcher & Berger 1972) with an aureole up to 1.5 km in width. Hornfelses mainly form the aureole and towards the contact they become increasingly disrupted and mobilised and this is thought to be due to the release of fluids generated by prograde dehydration reactions of pelite due to rising magma (Oglethorpe 1987). Localised overpressures result in fracturing of competent quartzites, with flow occurring in adjacent pelites as suggested by fragments of quartzite floating in the pelite matrix, (Oglethorpe 1987). Pitcher & Berger (1972), on the other hand believed the "mobilised" pelites were the result of partial melting of the pelite. The hornfelses are composed of biotite, andalusite, cordierite, with rare garnet, whilst fibrolite and robust sillimanite are common near the contact. The low abundance of staurolite is interpreted as being the result of rapid heating of the aureole by the intruding Thorr Granite, overstepping the staurolite reactions, (Naggar & Atherton 1970). The width of the fibrolite hornfelses is variable, being a few 10's of metres to up to 1-2 km in the Lettermacacward and Thorr areas. Away from the granite, biotite porphyroblasts and muscovite define a foliation which Rickard (1963) attributed to the result of the forceful intrusion of the pluton. The Cashel Belt of Superimposed folds (Rickard 1962) which were also believed to be the result of forceful emplacement of the Thorr pluton, were regarded by McErlean (1993) as regionally D_3 in age, using the chronology of Hutton (1983), forming before the intrusion of the Thorr pluton. Rickard (1963) also suggested that the Crockator strike swing was tightened to some degree by the intrusion of the Thorr Pluton, although the later intrusion of the Main Donegal Granite to the south also further tightened this structure.

With regards to the emplacement of the Thorr pluton, Pitcher & Berger (1972) envisaged localised uplift and compression associated with the rising granite allied to active stoping and reaction with the country rock as the main method of emplacement. The lack of accommodation structures in the aureole led the above authors to put emphasis on stoping, with the country-rock rafts in the southern part of the pluton representing arrested stoping caused by more rapid cooling of the magma due to the close proximity to the roof of the pluton.

McErlean (1993) put more emphasis on a tectonic model for the emplacement of the pluton, with initiation of pluton construction occurring in the south, (in the

Thorr area) relating to movements along the Main Donegal Granite Shear Zone, (MDGSZ) (before the intrusion of the Main Granite itself), with the countryrock architecture playing an important role in the siting of melts (see figure 3:4).

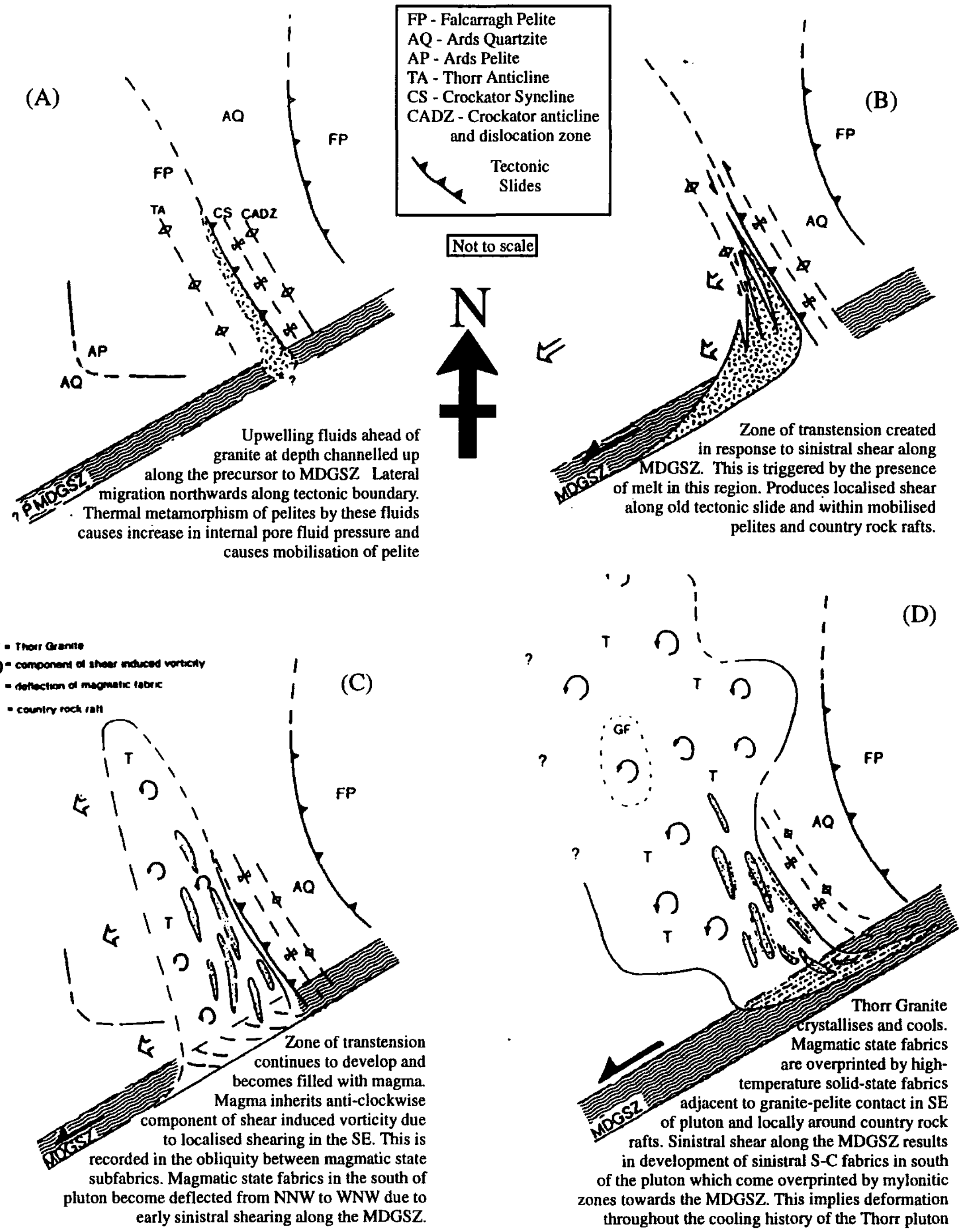


Figure 3:4:- The model for the emplacement of the Thorr pluton (*after* McErlean 1993)

The following model of emplacement is taken from McErlean (1993). Ascending magma batches channel up the MDGSZ exploiting the NNE trending anisotropies in the present Thorr area, (i.e. fold axes, tectonic slides and lithological contacts). Subsequent dehydration reactions in the pelites cause mobilisation of hornfelses. Granitic pulses forcefully intrude into the mobilised Upper Falcarragh pelites producing localised folds whilst the quartzite responds in a more competent manner showing top to the west displacements along pre-existing dislocations. Increasing magma input causes ENE sinistral reactivation of earlier structures accompanied by the development of a NNW trending transtensional cavity. The cavity continues to propagate northwards in an essentially passive manner whilst sinistral shear remains on the NE-SW shear zone in the south. Deformation related to emplacement produces magmatic fabrics (PFC fabrics of this thesis) which are broadly parallel to the contacts resulting from "buoyancy head" forces within the magma, although McErlean (1993) identifies a sub-fabric which shows a consistent component of sinistral shear throughout the pluton. This author attributes this to shear induced vorticity related to the Main Donegal Granite Shear Zone towards the south. In the latter area the magmatic fabrics become deflected by movements along the shear zone up to 1.5 km from the present day Main Donegal Granite. During the cooling of the pluton CPS and solid-state fabrics developed with sinistral S-C fabrics and zones of mylonites forming in southern areas of the pluton related to continued movement along the Main Donegal Granite Shear Zone. Hutton & Alsop (1996) have reported evidence of early sinistral shear on the N-S trending eastern contact of the Thorr pluton and have implied an emplacement model involving sinistral transtension on a shear zone related to the NNE-SSW lineament.

3:3:2 The Fanad Pluton

The Fanad granite is exposed on the two adjacent peninsulas of Fanad and Rosguill as fragments of a large pluton, of which most is now submarine (see figure 3:1). On Inishowen, (Tullagh Point) there is another part of an intrusion which Pitcher & Berger (1972) thought might belong to the Fanad Pluton, but geochemical analysis has shown that, whilst it is petrographically similar, it is a separate pluton, (O'Connor *et al.* 1987). The Fanad pluton is an isolated member of the Donegal Batholith and hence relative age relationships are more difficult to ascertain. The absolute age is 402 ±10 Ma, (Rb-Sr whole rock isochron), unequivocally relating the Fanad pluton to other members of the batholith (O'Connor *et al.* 1987). The pluton post-dates all the regional D_{1-3} movements as indicated by extensive country rock rafts within the granite which all record these earlier phases. These xenoliths belong to the higher formations of the Appin Group and display a cruder "ghost-stratigraphy" than that preserved in the Thorr Pluton, (Pitcher & Berger 1972). Deformation within the granite is weak, with mafic enclaves recording minimal to low strain. The fabric is broadly parallel to the contact, though in some areas there are local discordances to the host rock, suggesting the influence of weak tectonic stresses.

The Fanad pluton itself shows many similarities to the Thorr Pluton, though the former is more dioritic, to almost appinitic in character, (Pitcher & Berger 1972). On west Rosguill there is a moderately foliated monzodiorite, separated from an almost structureless granodiorite on NW Fanad by the Melmore Septum and associated migmatites. The granodiorite on Fanad outcrops as far as the Gortnatraw Septum, east of which a foliated tonalite reappears. Pitcher & Berger (1972) believed the Melmore migmatites are the same age as the rest of the pluton and represented a highly mobilised and injected remnant of a roof septa which dies out at depth. The typical pluton type, the monzodiorite, is composed of plagioclase, (An_{30-40}), quartz, K-feldspar, hornblende, biotite, zircon and apatite. As with the Thorr Pluton there is a contact facies, on average a few hundred metres wide, although a wider belt occurs between the Melmore and Gortnatraw Septums, and is of more granodioritic composition, differing from the host monzodiorite with less hornblende and increased K-feldspar and quartz. Within the aureole there is little evidence of any deformation apart from weak crenulations and the occasional upright folds, indicating that much of the aureole is relatively static (Pitcher & Berger 1972). In the area of Glinsk, on Fanad, the effects of intrusion are clearly superimposed on the regional structures in pelites for up to 1.75 Km away from the contact. On approaching the aureole progressively higher grade minerals are encountered, i.e. garnet, andalusite, fibrolite and cordierite, (accompanied by a decrease in muscovite towards the granite) and robust sillimanite appears up to 200 metres from the contact. Naggar & Atherton (1970) also noted the presence of corundum at some locations and attributed its presence to the rapid intrusion and heating of the aureole by the Fanad Pluton.

The absence of deformation within the both the granite and aureole implies a relatively passive nature to emplacement of the pluton. Pitcher & Berger (1972) stated the abundance of xenoliths and roof septa implied the pluton at the present level of exposure is granite is close to its roof. The same authors claim that most of the granite was emplaced by piecemeal stoping in the form of steep-sided apophyses that were united at depth.

3:3:3 The Ardara Pluton

The Ardara Pluton occupies the low-lying ground to the SW of the Main Donegal Granite and the area to the north of the town of Ardara. The pluton forms a broadly concentric outcrop with an average diameter of 8.5 km, apart from the elongated apophysis in the SE known as the "stalk" (Pitcher & Berger 1972). The stalk of the Ardara Pluton is cut by the Main Donegal Granite, and it is believed that the intrusion of the former pluton (and possibly the Toories Granite) followed the intrusion of the Thorr Pluton. A Rb-Sr (whole-rock isochron) age of 405 ±5 Myr has been obtained for the pluton (Halliday *et al.* 1980). Associated in space and time with the Ardara Granite are a whole series of appinitic bodies, of which the majority were intruded before the granite.

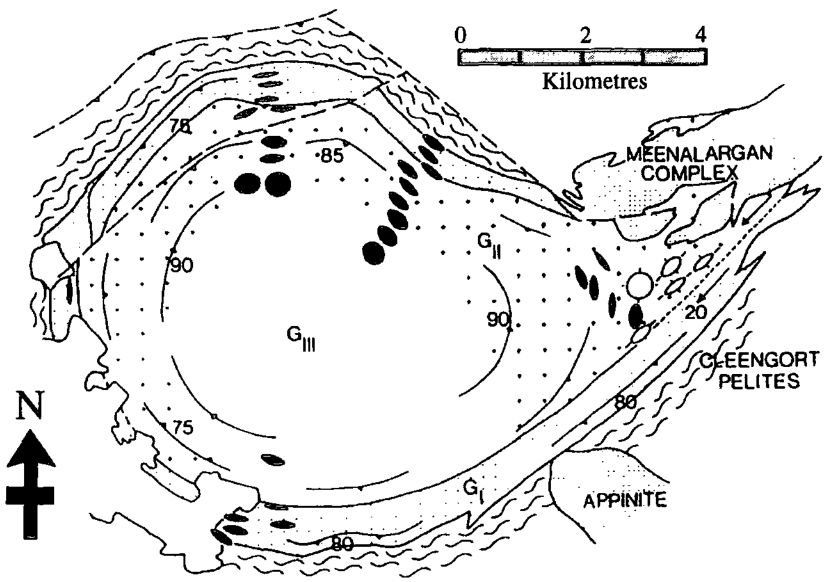


Figure 3:5:- The Ardara Pluton (black ellipses show strain variation) (*after* Holder 1979).

The pluton comprises three compositional units, intruded almost entirely into the Upper Falcarragh Pelites, the exception being where units cut the Meenalargan Appinite Complex (see figure 3:5). The outermost unit is a coarse-grained monzodiorite, G1, varying in width from 230-900 metres and forming the majority of the perimeter of the pluton, where it forms a smooth and sharp contact with the host meta-sediments. Xenoliths, (often mafic enclaves) within this unit are aligned and flattened into the foliation plane broadly parallel to the contact of the granite. Inwards from G1 is a megacrystic tonalite-granodiorite, G2 which forms a broad horse-shoe shaped outcrop, open to the south, occupying much of the stalk where it breaches the G1 body in this area, lying directly in contact with and incorporating xenoliths of the Meenalargan Complex (Holder 1979). The contact of G2 with G1 is everywhere sharp. Occupying the central area of the pluton is a coarse-grained equigranular granodiorite, G3, which has a sharp contact with G1 in the southern part of the pluton, but a gradational contact with G2. The three units area believed to be separate pulses and are not the result of in situ differentiation (Holder 1979).

Within the pluton the deformation increases towards the margin of the granite, as indicated by mafic enclaves and the development and intensification of a foliation, (equivalent to the S4 foliation in the country-rock (Meneilly 1982), (Akaad 1955; Pitcher & Berger 1972; Holder 1979), although within the "stalk" there is superimposed deformation related to the later Main Donegal Granite. Work by Molyneux & Hutton (in press) on mafic enclaves and Fry spacing analysis have confirmed that strains are pure flattening, (i.e. K<<1). The intensely deformed aureole has been described earlier, (see section 2:5:1:2) but is typified by concentric, upright schistosity which obliterates all earlier structures within a distance of 100-250 metres from the contact, (Meneilly 1982).

The emplacement of the Ardara Pluton has been the subject of extensive previous work with it having been interpreted as; an intrusive diapir, (Akaad 1955; Pitcher & Berger 1972); an expanding balloon, (Holder 1979); ballooning plus a large component of stoping, (Vernon & Patterson 1993). The most recent addition to this problem is by Molyneux and Hutton (*in press*) who present a model of oblique ballooning with the point of emplacement to the SW of the approximate geometric centre of the pluton. These authors propose a northward directed expansion with the aureole in these areas showing circumferential compression and stretching, whilst to the south there is a sinistral shear zone which they consider to be a splay of the MDGSZ. De-straining of the mafic enclaves within the granite, assuming ballooning, accounts for 80% of the area of the pluton (Molyneux & Hutton, *in press*).

3:3:4 The Donegal Appinites

Within Donegal there is a whole suite of boss-like intrusions which belong to a suite of igneous rocks known as appinites. The majority of these intrusions are located in the vicinity of the Ardara Pluton, though more isolated bodies have been documented elsewhere in the Donegal region. The appinites are believed to be associated with the earlier members of the Donegal Batholith, notably the Ardara Pluton, (Pitcher & Berger 1972). Dating of the Kilrean appinite has given an $^{40}Ar/^{39}Ar$ age of 410 ±6 Ma, (O'Connor *et al.* (1982). Cross-cutting relationships also support the slightly older age for the majority of the appinites, although Pitcher & Berger (1972) state there may be a slight overlap between the intrusion of the appinites and the emplacement of the Ardara granite.

Petrographically these bodies are diverse but are typified by hornblende (usually idiomorphic) in a groundmass of plagioclase and quartz, with rare to absent K-feldspar. Other mafic minerals, such as pyroxene or biotite may occasionally be present instead of hornblende. Compositionally appinites are extremely variable, but are usually basic to ultrabasic (Pitcher & Berger 1972). In all the Donegal appinites the grain size is coarse, even in the small intrusions, a property attributed to high pressure and volatile content, (e.g. H_2O) which aids diffusion of ions within a melt to crystal surface, (*in* Pitcher & Berger 1972).

The appinite intrusions greatly vary in size from 100 metres in diameter to 1500×500 metres and are often compositionally zoned. These bodies are commonly lenticular with steep to vertical contacts having exploited the structural grain of the country rock. As well as intrusive bodies there are agglomeratic bodies characterised by breccias and "pipe-like" structures implying that many of the appinites were emplaced at high pressures and velocities associated with devolatilization of magmas.

Appinites are generally believed to be K^+ and H_2O rich magmas which have been derived from mantle or lower crustal regions and are genetically associated with granitoid magmas (Pitcher 1993). The relatively high initial ε_{Nd} for the Ardara granite (Dempsey *et al.* 1990) also supports a mantle or lower crust origin reinforcing the appinite-granitoid affinity, which is a common association world-wide and is not just a feature of the British Caledonides (Pitcher 1993).

3:3:5 The Toories Pluton

Exposed on the southern part of Aranmore Island and on the small islands of Iniskeeragh, Illancrone and Roaninish are fragments of quartz monzodiorite-monzotonalite belonging to a dominantly submarine pluton, known as Toories, (Pitcher & Berger 1972). In all the areas of outcrop there is a steep foliation, broadly parallel to the contact, which changes orientation defining an arcuate boundary to the pluton. The flattening of mafic xenoliths within the plane of foliation suggests $k<<1$ as seen in the Ardara Pluton (Pitcher & Berger 1972). On Aranmore the pluton is in sharp contact with the Ards Quartzite and syn-plutonic deformation is demonstrable where associated pegmatites are isoclinally deformed by the granite. Despite the limited exposure the analogy is with the Ardara pluton where a mechanism of forceful intrusion, possibly by ballooning has occurred. No geochemical data or absolute ages exist for this body, although near the contact xenoliths of Thorr Granodiorite are visible within the granite whilst members of the Rosses porphry dyke swarm cross the Toories pluton (Pitcher & Berger 1972). This relates its intrusion in time close to that of the Ardara pluton.

3:3:6 The Rosses Pluton

The Rosses pluton forms a centred complex composed of four relatively homogeneous masses, plus two associated phases of minor intrusives. The pluton occupies the flat lying ground around, and to the north of the small town of Dungloe in an area known as the Rosses. The pluton is sub-circular with an average diameter of 8.5 km with the contacts consistently dipping steeply away from the centre (figure 3:6).

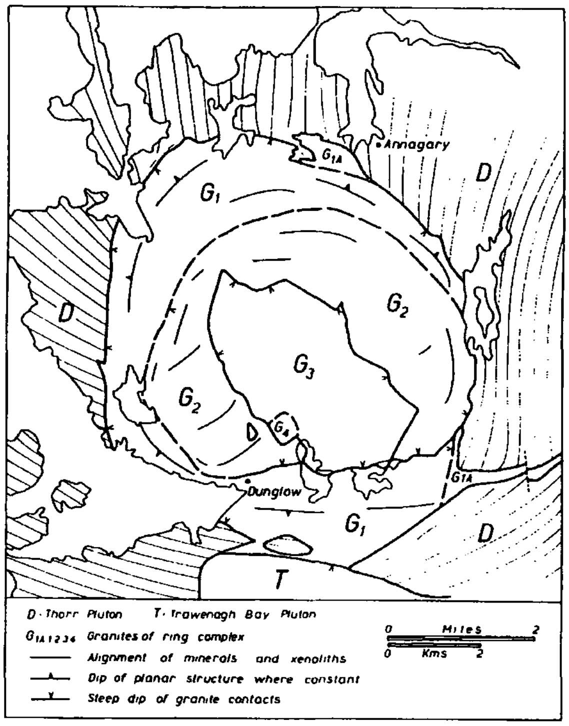


Figure 3:6
The Rosses centred complex (*after* Pitcher & Berger 1972).

The Rosses complex is intruded entirely into the Thorr Granodiorite although to the south it is truncated by the Trawenagh Bay Granite, clearly constraining its age compared to these other plutons. More equivocal evidence are the presence of certain dykes which are believed to represent members of the Rosses dyke swarm which cut both the Toories and the aureole of the Ardara Pluton, (Pitcher & Berger 1972). Thus the emplacement of the Rosses complex lies between the Ardara (+Toories) and the Trawenagh Bay-Main Donegal Granite. Halliday *et al.* (1980) obtained an age of 404 ±3 Ma, with an initial $^{87}Sr/^{86}Sr$ value of 0.7062 for the complex, broadly agreeing with the field evidence. The chronology of emplacement within the pluton is:-

1) A variable suite of microgranite sheets, trending NE-SW, broadly tangential to the pluton. The size of the sheets is highly variable, with the thickest, the Crovehy Sheet, being 210 metres thick.

2) All the above members are truncated by the G1 outer granite (see figure 3:6), which forms the outer granite of the complex. The contact with the Thorr Granodiorite is everywhere sharp, with no alteration of the former by the latter.

3) The contact between the medium grained G1 and the coarser grained G2 is transitional over a distance of approximately 100 metres. In the east of the pluton G2 breaches the earlier G1 to come into contact with the Thorr Granodiorite. This all

suggests G1 was not fully crystallised but was capable of being fractured and subsequently homogenising with G2.
4) A suite of porphyry dykes cuts the Thorr Granodiorite, G1 and G2 and displays sharp contacts and slight evidence of chilling.
5) The finer grained G3 cuts the earlier phases and the porphyry dykes and has everywhere sharp contacts, often polygonal on an outcrop scale. Along the contact between G2 and G3 there is an occasional narrow pegmatite fringe separating the two granites, (and G3 from G1 in the southern part of the pluton). Within certain parts of G3 there are muscovitic-rich areas that are the result of late replacement, i.e. greisenization, often gradational with the typical G3 facies.
6) The small body of muscovite-rich granite, G4, is a distinct intrusion is related to garnetiferous pegmatites and aplites, both of which cut G1-G3.

Pitcher & Berger (1972) report the virtual absence of any mineral alignment within the granites, apart from weakly developed structures in the G1 and G2 members, all of which are broadly parallel to the contacts, a feature which these authors attribute to a primary flow structure in the Cloosian sense.

The main biotite granite phases, G1-G4, are uniform in mode and texture, although there is a gradual decrease in biotite from 5% in G1 to 2.7% in G3. The muscovite containing varieties of G3 and G4 are typified by replacive muscovites enclosing relict feldspar. Furthermore the alteration of the G3 biotites has produced beryl-quartz-muscovite pegmatites, some of which contain appreciable concentrations of beryl. The microgranite dykes have a wide ranging composition with light and dark microgranites reflecting the amount of biotite they contain, whilst some of these sheets also contain muscovite.

The lack of deformation within the pluton and the adjacent granodiorite host strongly suggests passive emplacement by a cauldron subsidence mechanism, involving the sinking of a central block allowing magma to fill a void of progressively decreasing volumes (Pitcher & Berger 1972). G2 was intruded into a partially consolidated G1 which was capable of fracture and limited flow, whilst there seems to be greater time lapse between G2 and the intrusion of G3. Evidence for this is the N-S trending porphyry dykes which display chilled margins against the earlier phases, (G1 & G2) implying that G2 was totally consolidated. The truncation of the porphyry dykes and the polygonal nature to the contact is further evidence to support this thesis. Pitcher & Berger (1972) believed the porphyry dykes intruded into N-S fractures created by the uprising G3 pulse. During the late evolution of the pluton there was concentration of volatiles under pressure causing greisening and the formation of late muscovite.

3:3:7 The Barnesmore Pluton

This pluton occupies an area of 52 km^2 forming the mountainous Blue Stack range of central Donegal (see figure 3:7). The pluton lies to the south of the Leannan Fault, forming an outlying member of the Donegal Batholith, emplaced into psammites and pelites belonging to the Termon Pelite Fm.\Lough Eske Psammites\Lough Mourne Fm. of the Ballybofey Succession (Alsop & Hutton 1990). In relation to the other members of the batholith the age relationship is uncertain, apart from the rather equivocal evidence that breccia pipes similar to those seen around the Ardara Pluton intrude into the pluton. A Rb-Sr whole-rock isochron age of 397 ±7Myr has been established by O'Connor *et al.* (1987) with an initial $^{87}Sr/^{86}Sr$ of 0.7063, at least temporally relating it to the other granites in Donegal. The pluton has been dissected by a series of NE-SW trending faults which display dominantly sinistral offsets, plus an additional component of dip-slip, of which the Belshade Fault is the most prominent. This offsets the granite sinistrally by 3.5 km producing localised cataclasis and reddening of the feldspars within the adjacent granite.

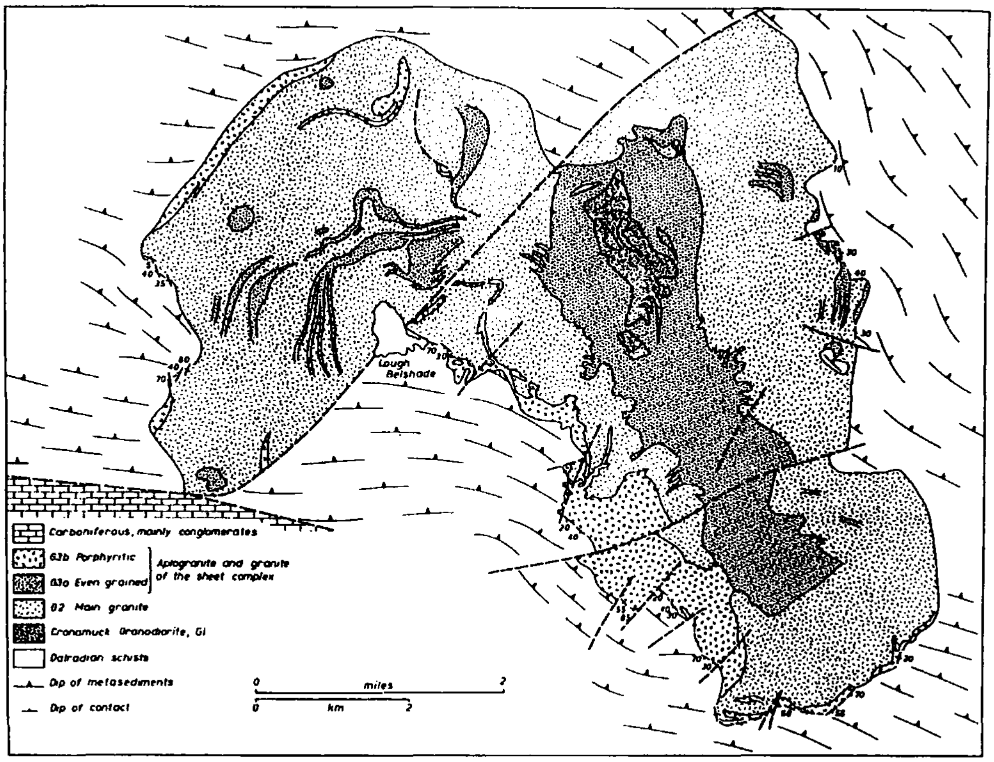


Figure 3:7:- The Barnesmore pluton (*after* Walker & Leedal 1954)

A noteworthy feature of this pluton is the totally undeformed envelope in which structures within the psammites show no deviation from the regional trend right up to the contact with the granite. The aureole of the granite is very narrow, typified by the appearance of unoriented andalusite approximately 200 metres from the contact, with the patchy development of hornfelses mainly consisting of biotite-muscovite schist containing augened relict oligoclase from the regional metamorphism (Pitcher & Berger 1972). The narrow aureole and the absence of deformational structures within both the country rock and granite implies passive emplacement at high level within the crust (~5Km). The country rocks were originally some of the deepest buried rocks in Donegal on the inverted limb of the Ballybofey Nappe. The age of D_3 deformation was dated between 460 Ma and 397 Ma implying uplift and erosion rates of 10-12 km within a period of 50 Ma preceding the emplacement of the Barnesmore pluton, (Alsop & Hutton 1993; *and references therein*).

The Barnesmore Pluton consists of three distinct phases which broadly have the same compositions but vary in both grain size and mode. Micas rarely make up 5% of the volume with plagioclase compositions in the cores being An_{20}. The order of emplacement is as follows:-

1) the Cronamuck Granodiorite, (G1):- a medium to fine-grained granodiorite forming a small outcrop near the centre of the pluton.

2) the Main Granite, (G2);- this intrudes the G1 phase and comprises the majority of the pluton at the present level of exposure. The granite is a very homogeneous coarse-grained, non-porphyritic leucocratic rock although near the margins of the pluton, there is very slight basification. G2 forms much of the outer contact of the pluton, with the attitude of the contact varying from nearly vertical to gently inclined and **always** dipping outwards from the pluton. Occasionally there are fragments of the roof preserved in the granite where the contacts are shallow, though xenoliths are rare to absent implying stoping was not an important process at this level of exposure.

3) the Sheet Complex, (G3):- consists of two facies which both intrude G1 and G2. An internal body, G3a is composed of a fine-grained aplogranite, plus associated layers of pegmatite, whilst the G3b granite is a porphyritic aplogranitic facies common near the margin of the pluton. The central mass of G3a has shallow angle sheet-like apophyses intruding into the G2 granite.

The absence of deformation within the aureole and in the pluton, suggests a passive mechanism of intrusion, whilst the lack of xenoliths casts doubts on stoping processes. The emplacement of G1 and the main body of G2 was permitted by cauldron subsidence of a domed block of country rock psammites along an outward dipping arcuate fracture (Walker & Leedal 1954; *in* Pitcher & Berger 1972). The presence of appinitic type breccia pipes containing unaltered schists which have been brought up from depth either represent material from the floor of the pluton or part of the sunken countryrock block. Further evidence for cauldron subsidence is shown by the emplacement of G3 into a void produced by the sinking of the central areas of G1 and G2. In the south the central G3a forms a large sheet which dips steeply to the NE.

Walker & Leedal, (1954) observed this sheet to link up with a thick sheet of G3b which dips to the SW in a feature they described as an "igneous arch". The G1 granodiorite in the centre of the pluton is where the G3 arch has been eroded to expose the partially sunken block of G1 and G2 underneath. The unsupported weight of the remaining G2 causes the development of broadly concentric fractures which allow G3 to intrude in sheet-like bodies along these fractures.

3:4 The Main Donegal Granite

3:4:1 Introduction

The Main Donegal Pluton is the largest, presently exposed member of the Donegal Batholith, extending from the Gweebarra Estuary in the SW, to the small village of Glen, some 50 Km towards the NE (see figure 4:1 and Map D of this thesis for location names). The pluton obtains a maximum width of 11 Km, having an overall area of 450 Km^2, including the weakly deformed but petrographically similar Trawenagh Bay Granite. The mountainous areas of the Derryveagh and the Glendowan provide excellent glaciated surfaces in which to study this granite. The following account may seem somewhat exhaustive, but the author feels that it is important to have this background information, as it allows a more greater understanding of the detailed fieldwork which ensues in the next three chapters.

The Main Donegal and Trawenagh Bay Granites form the youngest members of the Donegal Granite complex, with the former containing xenoliths of both the Thorr and Ardara plutons. The G1 member and the porphyry dykes of the Rosses pluton are truncated by the Trawenagh Bay Granite (Pitcher & Berger 1972). Despite being the youngest member, the Main Donegal Granite is typified as being a highly deformed, elongate pluton, broadly concordant to the envelope rocks apart from the highly discordant margin in the Maas area, (Iyengar *et al.* 1954). The intensely deformed aureole contains kyanite and sillimanite bearing schists, more usually associated with regional metamorphism, rather than contact hornfelses as seen around the other plutons in Donegal (Pitcher & Berger 1972). Within the pluton there is granitic banding and abundant, preferentially aligned country-rock xenoliths. These two features are parallel to the strong NE-SW trending, sub-vertical mineral alignment ubiquitous to the granite and the adjacent envelope. The Trawenagh Bay Granite is petrographically and geochemically similar to the Main Granite, although it is distinguished by a much lower degree of deformation. In the western part of the Main Granite, over a distance of 0.5-1.0 km, the strong fabric disappears, as does the banding and the xenoliths lose their preferred orientation and generally become less abundant, (Pitcher & Berger 1972). Furthermore the envelope of the Trawenagh Bay

Granite is totally undeformed and shows no evidence of forceful intrusion, being a narrow zone of static hornfelses.

Although discussed and described as a poorly distinguished part of the Donegal granites as a whole by earlier geologists, it was not until the work of Pitcher & Read (1959) that this pluton (Main Donegal Granite) was fully distinguished and named. The paper contained a map of the pluton, summarising the work of Tozer, Pande & Cheeseman, plus extensive mapping by the authors themselves. The important features of this paper will be now summarised.

3:4:2 The main features of the Pitcher & Read (1959) model

The Main Donegal Granite is a highly variable pluton composed of biotite, K-feldspar, plagioclase and quartz with muscovite, apatite and epidote as the most common accessories. There is a broad distinction between a medium-grained, equigranular granodiorite in the south-east, whilst a coarser grained porphyritic granite, with more variable biotite content and higher proportion of microcline, predominates in the NW (see figure 3:8). Despite this distribution, there are situations where coarse microcline-rich belts occur in the southern facies, whilst portions of biotite granodiorite form rafts in the NW facies implying that the porphyritic granites of the NW are younger. Near the SW contact of the granite in the Derkbeg-Straboy area, and northwards in the Galwollie Hill-Croaghleconnel area there are abundant pegmatites, which were interpreted as expressing roof phenomena (Iyengar *et al.* 1954), although in the latter area the presence of the Trawenagh Bay Granite complicates these relationships.

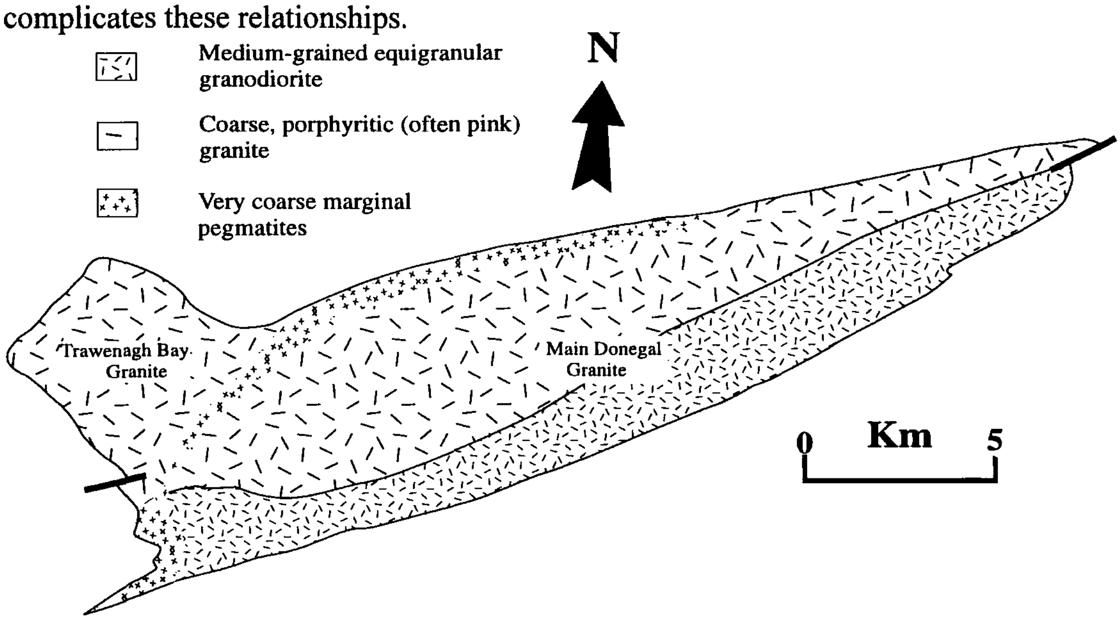


Figure 3:8:- The main petrographic variation within the Main Donegal Granite (*after* Pitcher & Read 1959).

Within the Main Donegal Granite there is a mineral alignment, crudely developed in the central areas but strengthening towards the margins to form a strong sub-vertical, foliation consistently oriented NE-SW with an associated sub-horizontal lineation (see figure 3:9). Pitcher & Read (1959) believed this mineral alignment was due to flow of the magma, although at the margins this alignment intensifies due to marginal solid-state shearing of the granite.

The banding within the granite is usually parallel to the mineral alignment apart from local discordances. The banding is due to three varying parameters within the granite; grain size, biotite and microcline content, with fine, medium and coarse bands observable with the latter two generally more common. There are several types of banding within the pluton although the most common type is regular banding. In this, lighter, coarser bands alternate with medium-grained granodiorite, the latter of which resembles the southern facies. These bands can be laterally very consistent although occasionally the coarse bands can be observed to enclose or cut the darker bands, implying the latter is older. Pitcher & Read (1959) observed the banding to be often localised around raft-like masses of fine-grained granite, e.g. the Doocharry Synform, where the fine-grained bands become more common towards the core. On the flanks of the core the banding is regular whilst at the nose it is irregular. The above authors interpreted this relationship as a streaming out of less viscous granite around earlier formed granite; "*...a flow structure in heterogeneous material of different consolidation points*". In summary the mineral alignment and banding were the structures produced by primary flow.

The contact of the pluton is everywhere sharp, although this is complicated by the fact that abundant granite sheets in these marginal areas form a zone up to 800 metres wide. The width of the sheets varies from small veins to sheets several hundred metres thick with the concentration increasing towards the contact until they eventually coalesce to form the main body of granite. The composition of these sheets is highly variable ranging from granodiorite and granite, typical of the main body, to aplite-pegmatite, the latter of which are often garnet-bearing. Mostly the sheets have lenticular geometries having exploited the schistosity of the country rocks, although where intruded into quartzite they are usually thicker and more regular in width - a *lit-par-lit* style of intrusion.

The sheeting in the country-rock gave Pitcher & Read (1959) a clue into the understanding of one of the more remarkable features of the Main Donegal Granite; the "raft-trains". These zones of abundant xenoliths, assimilated to varying degrees, although still recognisable as country-rock material, i.e. Dalradian meta-sediment and meta-basite, plus xenoliths from older plutons. Generally the quartzites are coarsened due to recrystallisation; pelitic rocks contain sillimanite, often transformed to muscovite; calcareous rocks form either calc-silicates or skarns and metadolerites form coarse-grained plagioclase amphibolites which contain biotite.

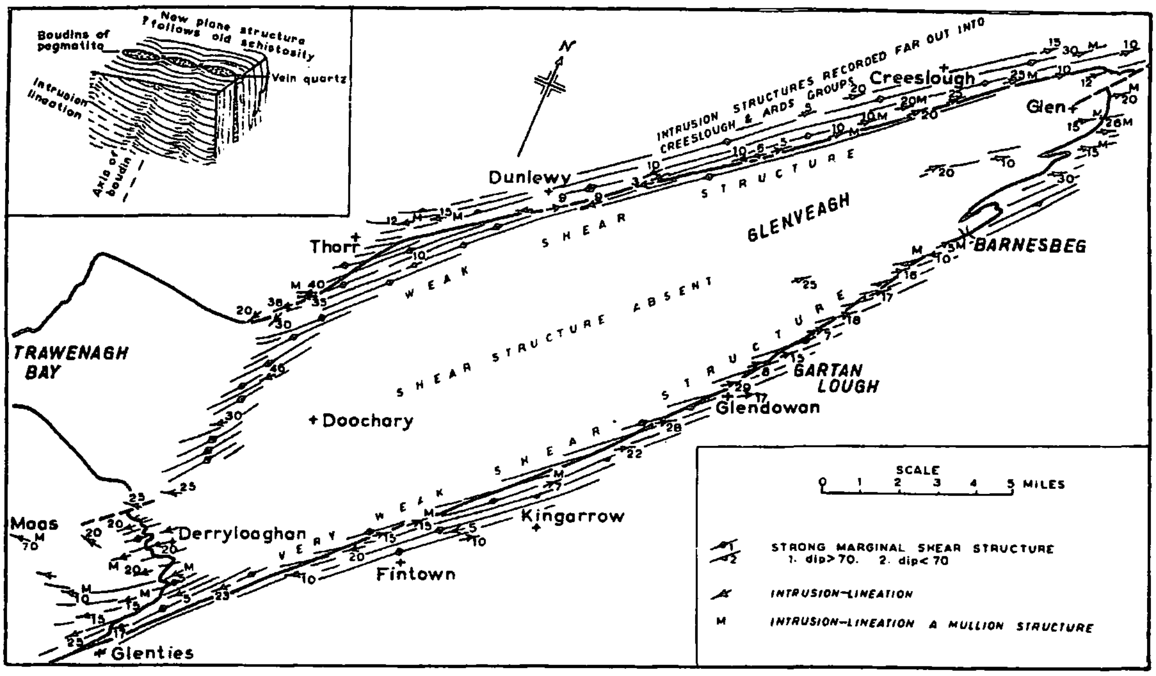


Figure 3:9:- The marginal deformation within the Main Donegal Granite (*after* Pitcher & Read 1959)

Often these xenoliths occur in groups of either mixed or single lithologies forming linear trains, termed "raft-trains", within the granite. These raft-trains are almost always parallel or sub-parallel to the mineral alignment and banding, implying their distribution had also been controlled by flow of the magma (Pitcher & Read 1959). The authors stressed that the individual rafts were "free-swimming" in the granite and they were not roof pendants, but as a whole the raft-trains could be shown to belong to either the roof or the wall-rocks of the pluton. In the SW part of the pluton, the distribution of the raft-trains allowed the authors to map the stratigraphy in some detail, between Fintown and Maas. Figure 3:10 shows the distribution of the raft-trains within the pluton, of which six major zones can be identified. In describing each of these zones the terminology of Pitcher & Read (1959) will be used.

The Border Zone (1):- occupies the NW length of the pluton from Cock's Heath Hill to Brockagh, a distance of 38 km, forming a zone a few hundred metres wide. The majority of the rafts are blocky and streaked-out xenoliths of Thorr Granodiorite, though amongst the granodiorite there are occasional pelite and quartzite xenoliths. In the SW part of the zone the Thorr granodiorite-diorite is only preserved as large para-autochthonous masses ranging from a few 10's of cm's to a large mass up to 2.5 x 0.25

km in size, appearing to be plucked of the walls by later granite apophyses. The host granite is a pale, often banded granite, associated with streaky pegmatites, clearly observable on the SW side of the Poisoned Glen. Towards the NE the zone narrows from 400-500 metres in the Dunlewy area to 200 metres in the Kingarrow, (near Callaber Bridge), beyond which the zone diminishes into a narrow strip of granite containing highly streaked-out Thorr xenoliths, resembling "fluxion gneisses".

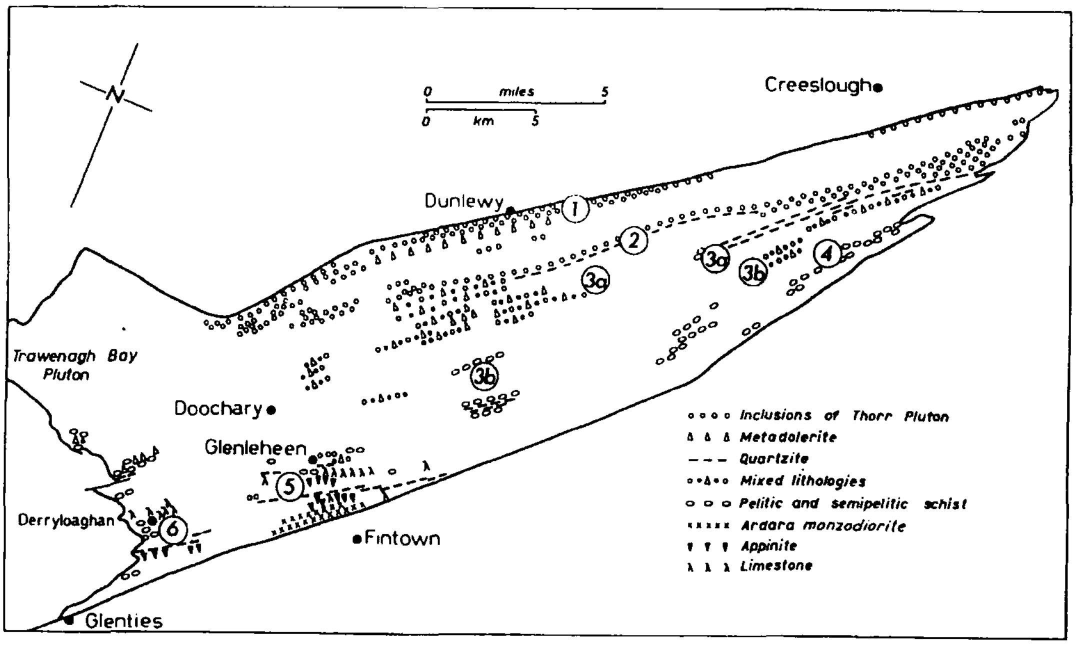


Figure 3:10:- The distribution of raft-trains within the Main Donegal Granite (*after* Pitcher & Read 1959). *(Numbers refer to text).*

The Derryveagh Zone (2a):- this is the longest of the raft-zones which traverses the well exposed plateau forming the Derryveagh Mountains, and maintains its identity from Bunbin Hill, near Glen, south-westwards to the Slieve Snacht area, a distance of 22 km. The rafts are composed of mainly angular xenoliths of Thorr Granodiorite and flaggy quartzites with occasional to rare meta-dolerites and pelites. On either side of the raft-train there are banded granites, a feature particularly well developed at the NW end of the Sruhanavarnis valley. The contacts of the xenoliths are generally sharp indicating very little interaction with the host magma, although smaller xenoliths tend to be strongly streaked out and flattened sub-parallel to the foliation. In the Glen Lough area, the xenoliths of Thorr Granodiorite are very large and angular, although they die out towards the base of Bunbin Hill (Pitcher & Read 1959). At this NE end the quartzite rafts are more numerous and appear to be "rooted" into the Ards Quartzite at Glenieraragh. Where the trend of the raft zone is oblique to the foliation, i.e. NE-SW, the raft- trains display an en echelon pattern with the individual rafts remaining parallel to the foliation e.g. to the south of Kingarrow.

The central raft-zones of Glenveagh (3a) and Binaniller (3b):- in the Crockmore-Derriscligh area all of the raft-zones are closely spaced, although towards the SW they progressively splay out, notably around Lough Veagh where the quartzite rafts split into two following opposite sides of the Lough. The southerly part forms a distinctly new zone, the 3a Glenveagh Zone, which witnesses a rapid decrease in the quartzite component towards the SW, becoming mainly pelite, meta-dolerite and calc-silicate with no xenoliths of Thorr Granodiorite. On average the zone is 400-500 metres wide, tending to be more discontinuous than the Derryveagh Zone, and forming closely spaced, but lithologically separate rafts. In the deep, drift-filled valley of Glenveagh the rafts are not visible, but along strike in the Sruhanavarnis area the zone re-appears on well exposed glacial pavements continuing to the SW in the mountains NW of Lough Barra and the small hamlet of Commeen. In the central portions of the pluton the pelite rafts are much more mobilised and feldspathised by the granite with localised assimilation having occurred. In the south-westerly extremes of this raft-zone xenoliths of Thorr Granodiorite re-appear as composite rafts of Thorr, pelite and meta-dolerite, implying the presence here of inclusion-rich Thorr Pluton prior to the intrusion of the Main Donegal Granite.

The Binaniller, (3b) zone takes its name from the a prominent hill visible in the Barnes Gap which has a major break of slope on its NW side at the base of which is the raft zone composed of dominantly pelite, metadolerite and limestone. Within the rail-cutting half-way along the Barnes Gap this raft-zone is very well displayed, steeply inclined towards the SE. The width of the zone is ~160 metres and is surrounded by relatively homogeneous granite, although amongst the rafts the granite is more variable with aplite patches being present. The Binaniller zone cannot be traced into the country-rocks as it is separated by 360 metres of homogeneous granite. Towards the SW the raft-zone can be traced for 10 km, forming a series of discontinuous trains, clearly separate from the Glenveagh zone, with the zones becoming progressively more separated by granite in this direction.

More isolated raft-trains are found in the central areas of the pluton and roughly lie along strike from the Glenveagh and Binaniller zones. Examples of these are seen on the SW slopes of Moylenanav and Crockastoller and are all heavily permeated by the granite. As Pitcher & Read (1959) stated these raft-zones appear to behave like "linear streamers" within the granite, being vertically discontinuous. Further to the SW of Moylenanav there is another raft-zone in the Gweebarra valley composed of pelite, meta-dolerite and also xenoliths of Thorr Granodiorite.

The Crockmore Rafts and septum (4):- around the summit of Crockmore there is a shallow dipping septum, composed of pelite with some meta-dolerite. The septum progressively gets narrower towards the SW, until near the N56 road, (around Lough

Acrobane) it appears to break up into a series of rafts, whose orientation becomes modified by "flow" within the granite. To either side of the granite are two homogeneous granites which are separated from each other by the septum. Along strike from the Crockmore Septum are a series of dominantly pelitic raft-trains, exposed in the small townlands of Losset and Whitehill. Even further to the SW on the southern flanks of Leahanmore and Croaghacormick there are three raft-trains resembling the material within the Crockmore Septum.

The Glenleheen Raft-Zone (5):- this forms a ~3 km wide zone of lithologically diverse trains which, when grouped as a whole, is traceable for ten or so kilometres. This raft-zone does not lie along strike from the other previously discussed zones whose trains shows a strong correlation with the Mass country-rocks. On a traverse from the SE to NW along the Fintown-Doocharry road, the first rafts to be encountered are angular xenoliths of monzodiorite, contained within fine-grained granite and pegmatite. Although highly deformed Pitcher & Read (1959) related these xenoliths to the Ardara Pluton, whose main outcrop is exposed further to the SW. These darker mafic xenoliths within a pale host are clearly visible from the road through the Carbat Gap.

The tonalites are succeeded by a 5 km long set of quartzite raft-trains, visible at the NW end of the Carbat Gap and also along the Lough Errig road to the SW. Within the small valley of Glenleheen raft-trains are rare within a medium-grained biotite granite. Instead there are large blocks of limestone and appinitic diorites resembling members of the Meenalargan Complex. Near Gubbin Hill, a large enclave of Falcarragh limestone (Pitcher & Shackleton 1966) has been quarried out. Along the side of the Lough Errig road large blocks of dark appinite are clearly visible, differing from the meta-dolerites by their coarser grain size and their distinctive weathering patterns, i.e. a type of carious weathering.

The most prominent member of this raft-zone is another quartzite belt around 270 metres wide and traceable for almost 13 km from Lough Muck to Glenleheen. Rafts up to 360 metres long display an en echelon geometry. To the NW of this prominent raft-train there are further trains of appinite forming a well-aligned set of inclusions. Beyond this is a set of limestone inclusions which have been quarried out beside the old Glenleheen School, which form a long train up to 7 km, clearly visible on the flanks of Croaghleheen. Finally in the NW there is a mixed zone of pelite, quartzite and meta-dolerite, including the Thorr Granodiorite. The quartzites here show considerable signs of reaction with the granite with the production of quartzified granite (Pitcher & Read 1959). This raft-zone is vertical and was interpreted by the above authors as wall rocks "spalled off" by the intruding granite.

The Derryloaghan Zone (6):- this is another distinct zone of diverse raft-trains here present along the SW contact of the Main Donegal Granite. The presence of highly pegmatitic granite around this zone is interpreted as a location lying close to the roof of the pluton, (Pitcher & Read 1959). At the SE end of the zone there are abundant xenoliths of appinite which clearly belong to the Meenalargan Complex, extending 2 Km into the pluton towards the Gaffaretmoyle area. To the NW of the appinite rafts is the longest raft-train of Sessiagh-Clonmass quartzites (the Mulnamin Siliceous Flags of Iyengar *et al.* (1954)) which is clearly traceable from the contact for 4 Km into the pluton. NW of this raft-train there are a series of disjointed rafts composed of calc-pelite, meta-dolerite, pelite and limestone, the latter of which often have thick skarns surrounding them. In comparison with the other zones, the Derryloaghan zone is more sporadic and consists of very large angular blocks.

From the distribution of all the raft-trains within the Main Donegal Granite, Pitcher & Read (1959) made the following conclusions:-

- The enclaves are isolated pieces within a granite and are not roof-pendants or screens separating different types of granite.
- The Border Zone, (1) is interpreted as a smear-zone generated by viscous flow within the granite in an essentially horizontal direction, with flow from the NE towards the SW.
- Lamellar, horizontal flow was also attributed to the consistent length of the Derryveagh Zone, (2).
- Zones (3a), (3b) & (4) can all be rooted into the roof, with the south-westerly splaying out of the rafts again due to flow from the NE to the SW.
- Zone (5) is composed of rafts derived from the wall-rocks, as indicated by their slightly oblique attitude to the contact, with flow towards the SW.
- The Derryloaghan Zone, (6) is the result of south-westerly flow beneath the roof of the pluton.
- the enclaves did not represent a true "ghost-stratigraphy", but instead were fragments incorporated from the roof and walls and conveyed from the NE to the SW by the magma. The pendants such as Crockmore were parts of the country-rock which had survived the forceful wedging due to the intrusion of large granite sheets.

Evidence cited for the forceful emplacement of the granite was the strong mineral alignment, banding and the intense structures in the marginal granite and immediately adjacent envelope. Towards the margin the mineral alignment becomes stronger with granite sheets becoming intensely sheared and boudinaged on axes normal to a sub-horizontal mineral stretching lineation, which Pitcher & Read (1959) called their "intrusion lineation", created by lateral flow of the magma rasping against its host, causing cataclasis of the cooling granitic material. In the NE end of the pluton this lineation plunges towards the NE-NNE, whilst in the SW part of the granite, the same lineation plunges to the SW-SSW.

Within the envelope of the pluton, Pitcher & Read (1960) believed the forceful intrusion of the granite generated new folds superimposed onto earlier regional structures. Away from the granite, small folds and an associated crenulation cleavage record the earliest development of structures related to emplacement. Towards the granite these folds become larger and progressively tighten with the formation of a strong axial planar schistosity adjacent to the pluton, with the formation of fold mullions in quartzites. The plunge of the mullions is coincident with the intrusion lineation (Pitcher & Read 1960). The same authors reported the notable increase in metamorphic grade towards the pluton from regional garnet chlorite schists (mid-upper greenschist facies) to coarse schists containing andalusite, kyanite, staurolite, sillimanite, garnet and plagioclase, (mid-amphibolite facies assemblages). Textural studies of these porphyroblasts attested to a syn-deformational origin and the forceful intrusion of the granite. The presence of the three alumino-silicate polymorphs not displaying replacement relationships to one another was interpreted as P-T conditions close to the triple-point, though Pitcher & Read (1960) did put some emphasis on the compositional controls of the envelope rocks.

The emplacement model for the Main Donegal Granite by Pitcher & Read (1959) was by lateral magmatic wedging and horizontal stretching, with the intrusion of granite in the NE and flow of a viscous magma to the south within a steep-sided pluton with a "pronged roof". The above model agreed with the orientation of the banding, the splaying out of the raft-trains, the intrusion lineation and the mineral alignment. Deformation was intense along the margin due to a higher viscosity contrast between the envelope and the crystallising magma. Pitcher & Read (1959) stated that if the intrusion lineation correlated with the viscous flow "*then the magma rose gently from the NE, undulated more or less horizontally towards the SW, and dived fairly steeply beneath what is now the south-western roof*". Therefore the mountainous areas of the pluton, i.e. Derryveagh and Glendowan, display the deeper portions of the pluton, as indicated by the more heavily permeated enclaves in these areas.

In the discussion to the Pitcher & Read (1959) paper Tozer and Harland, considered the emplacement of the Main Donegal Granite to be mainly by vertical ascent of granitic material with subsequent arresting of ascent causing axial extension of the pluton, possibly related to the confining pressures of the walls. Another problem that they addressed was how could horizontally moving magma produce folds in the envelope parallel to the direction of movement, even despite the authors claim that the granite tightened folds of earlier regional age. Hence Pitcher & Read's model of emplacement, whilst satisfactorily explaining many of the features of this pluton and its envelope rocks, did not, account for all of them.

3:4:3 A structural re-evaluation by Berger (1967) (*in* Pitcher & Berger (1972)

An attempt to correlate the deformation in the pluton (i.e. the banding and the mineral alignment), with the structures in the aureole, formed the basis of Berger's later studies (Berger 1967, unpublished thesis), much of this which was summarised in Pitcher & Berger (1972). These authors found that the mineral alignment cross-cut internal petrographic contacts, including banding and hence could not have been formed by classic magmatic flow as Pitcher & Read (1959) had originally believed. Pitcher & Berger (1972) favoured vertical ascent of granite sheets rather than horizontal wedging. The deformation within the pluton was interpreted as being the result of a NW-SE directed compression acting upon the pluton, during the cooling history with the greatest extension direction being sub-horizontal (Berger 1967). The origin of the banding within the granite was investigated by Berger (1971); Pitcher & Berger (1972) from which the following account is taken. Several types of banding exist within the pluton:-

i) Along the NW margin, the intrusion of a series of parallel, dark, microgranite dykes produces a crude banding.

ii) A more irregular type of banding results from NE-SW aligned, smeared-out xenoliths of pelite and earlier fine-grained granite. Often the former of these inclusions tend to be locally assimilated, producing lighter and coarser pegmatite patches between biotitic schlieren.

iii) In some areas the regular bands and possibly some of the other types appear to have been subsequently deformed, i.e. streaked out and folded. Pitcher & Berger (1972) termed these types of granite as showing "discontinuous banding" and characteristic in the marginal zones of the plutons where the strain is more intense.

iv) The more common type of banding is the regular banding described by Pitcher & Read (1959) which is most common in the north-western part of the pluton. The banding consists of coarser grained, light-coloured granite and dark, finer grained trondhjemite, almost always parallel (and steeply dipping) to the long axis of the pluton. Bands are variable but generally less than one metre in width, traceable along strike for up to thirty metres. Often the dark bands lens out within a host of lighter band with whole zones of banding fading out along strike into a host granite strongly resembling the coarser, lighter bands. In outcrop the contacts between the different bands are sharp, (though transitional contacts are seen), but in thin section the contact is not obvious due to interlocking of crystals, with a general increase in microcline feldspar moving from dark into light bands. Berger (1971) agreed with the view of Pitcher & Read (1959) that the darker bands are of earlier origin. Berger (1971) observed cross-cutting bands (which he termed "cross-bands") which were generally similar in composition to the light bands, though differing very slightly in texture. The cross-bands trend NE-SW and are usually common (or at least distinguishable) in areas where the host bands are discordant to the main trend, e.g. in the Doocharry area. The contacts of the cross-bands with their host is usually sharp but, like the host bands, transitional boundaries are seen. The cross bands displace the host regular bands to varying amounts, but on most occasions it is impossible to match regular bands across them, implying that the cross bands are planar zones of localised movement within the granite. No cross bands of just dark band material have been observed, although in some of the wider cross bands the darker bands produce discontinuous streaks parallel to the margins of the bands. The similarity in texture of the cross bands and the light host bands, plus the fact that dark bands never form cross bands on their own, also fuelled the train of thought that the darker bands are older (Berger 1971; Pitcher & Berger 1972). Therefore the cross bands are zones of movement after the host banded granite had solidified, with the abrupt truncation and lack of deflection of the host bands by the cross bands. Berger (1971) believed the host light bands to be formed in a similar fashion.

The major difference in petrography between the two sets of bands is the microcline content. In the light bands the microclines, which are up to 7 mm in diameter, give the light bands their light colour and apparent coarser appearance, whilst the dark bands contain little to no microcline (Berger 1971; Pitcher & Berger 1972). Often the lighter bands contain more quartz and lower proportions of biotite, though there is very little difference in content or abundance of the accessory minerals between both band types. The biotites in close proximity to the microclines are often chloritised in the light bands and myrmekite and anti-perthitic intergrowths may occur within plagioclases. Otherwise the main textural and modal differences are just due to the addition of microcline within the light bands. Within large microcline crystals smaller grains of plagioclase, quartz and biotite are arranged in a zonal fashion, strongly implying the relatively late stage growth of microcline within the banded granites (Pitcher & Berger 1972).

Therefore the petrography and textures of the bands supports the late stage addition of microcline to a predominantly trondhjemitic parent mush composed of quartz, plagioclase and biotite. The composition of the cross bands led Berger (1971) to suggest a model where potassium-rich fluids, (plus some mobile silica) were concentrated along localised zones of movement. The pluton was not totally consolidated due to the lack of cataclasis associated with banding, although the granite was capable of fracture, as supported by the sharp contacts of the cross bands. The process is similar in some ways to "filter-pressing" where " *shearing of a crystalline mush, particularly when it is still weakly knit together*" may result in fractures that "*instantly fill with liquid, (in this case potassium-rich fluids) from the interstices of the mesh and repetition of the action may give rise to banding, properly oriented with respect to the shearing stress*" (Pitcher & Berger (1972), p. 219 and references therein). The presence of early-healed shears, where dark bands have been dragged out implies that some areas of the granite were subject to lower strain rates allowing ductile deformation to occur, (Berger 1971). The source of the potassic-rich fluids was uncertain but the author suggested that the whole granite originally had a bulk granodioritic composition but the squeezing out of the potassic-rich fraction formed a granitic *(sensu stricto)* portion leaving behind a trondhjemitic residue. Because the unbanded areas of granites resembled the lighter bands Berger (1971) suggested that the parent granodiorite had been totally potassic-feldspathised, with all evidence of the darker bands being totally obliterated in these latter areas. Therefore, the areas of banded granites had only been partially converted by this process.

As described earlier there is a ubiquitous NE-SW trending foliation, which is most intensely developed in the marginal areas becoming weaker towards the centre of the pluton. Generally this foliation is vertical to sub-vertical, although on the SW margin at the NE end of the pluton, the foliation is gently inclined to the SE, (Pitcher & Berger 1972). The fabric is developed by the alignment of quartz, biotite and occasionally microcline. In the central areas of the pluton the fabric is an S-type, and moving towards the margin the fabric develops a linear component, i.e. L=S and L>S with X of the strain ellipsoid being sub-horizontal (the "intrusion-lineation" of Pitcher & Read 1959), (Pitcher & Berger 1972). Apart from certain late dykes and later localised shear zones the mineral alignment is penetrative in all units within the pluton. At Doocharry where the banding is discordant from its usual trend, due to shearing of granite around earlier granite phases the cross-cutting nature of the foliation is obvious. The foliation appears to be axial planar to the synformal arch of banding in the Doocharry area. The foliation, as well as affecting the granite is also present in the immediately adjacent country-rock, clearly indicating that mineral alignment is superimposed on the granite, i.e. an external tectonic force. Not all of the minor intrusions within the Main Donegal Granite have been affected by the deformation, the presence or absence of a mineral alignment allowing Pitcher & Berger (1972) to erect a chronology of the dyke phases.

The minor intrusives of the Main Donegal Granite include microgranites, pegmatites, felsites and late veins, all of which post-date the banding. In respect to the formation of the main foliation within the pluton some of the intrusives possess the NE-SW alignment whilst others have an oblique mineral foliation and finally some of these intrusives are totally undeformed (Pitcher & Berger 1972). A very broad chronology based on cross-cutting relationships suggest the microgranites are older than the felsites, whilst the pegmatites and aplites range in age from pre-microgranites to post felsites. The late veins are definitely the youngest intrusives genetically related to the Main Donegal Granite (Pitcher & Berger 1972)

➊ ***Microgranites***:- within the pluton there is a two-fold division of these intrusives based on composition and relative ages. The earlier microgranites tend to be more mafic and finer grained than the later ones (Pitcher & Berger 1972). In most cases later intrusions and protracted deformation have disrupted these early dykes into smaller pieces which nevertheless appear to retain their original trend. Many examples of these disrupted dykes are seen on the well exposed glacial pavements in the NW of the pluton, e.g. the Sruhanavarnis and Crobane Hill. The mineral alignment is parallel to the host foliation, though in some dykes the alignment is oblique to both the main trend and the margins of these dykes, with the latter indicating movements of the dyke walls relative to one another, (Pitcher & Berger 1972). Pitcher & Read (1960) believed these dykes were emplaced into embryonic fractures within a crystallising granite which was still capable of some ductile deformation which subsequently deformed and disrupted the dyke.

The later microgranites dykes are much more common, tending to be slightly coarser and lighter coloured. The dykes are often very regular in dimensions, showing no signs of deformation apart from where they cross particular meta-sedimentary enclaves, (Pitcher & Berger 1972). Again the foliation may be parallel to that of the host granite or oblique to the dyke walls and main foliation trend.

The contacts of the microgranite dykes with the host granite are always interlocking in both hand specimen and thin section. Petrographically the microgranites range in composition from granite to tonalites, although the earlier microgranites tend to have the more intermediate compositions, (Pitcher & Read 1960; Pitcher & Berger 1972).

➋ ***Felsites***:- these dykes are typically composed of fine to medium-grained granodiorites which may be aphyric or porphyritic. Often they show a grain size reduction towards the margin, possibly relating to a chilling effect or the result of magmatic flow sorting in the dykes, (Pitcher & Berger 1972). The authors report that most of these dykes are found in the south-western and western portions of the pluton, and show strong similarities in texture and mineralogy to members of the Rosses porphyry dyke swarm, although spatially the link is lacking. The dykes are characterised as being quite regular, though sometimes branching, having planar walls and traceable for a few hundred metres along strike. Often these dykes have been internally deformed with almost all of them having oblique foliations to both the dyke walls and that of the host granite, implying they have been zones of movement, as indicated by the difficulty in tracing pre-dyke markers between dyke walls (Pitcher & Berger 1972). The absence of cataclasis (apart from the felsites in the marginal areas) within these dykes strongly suggests that they were being deformed during and after emplacement but before their complete cooling. Within thin felsites the mineral alignment is often parallel to the to the walls of the dyke, whilst in the thicker dykes the foliations are more sigmoidal; "Z" shaped foliations indicating sinistral movements of the walls and "S" shaped foliations recording dextral offsets (Pitcher & Berger 1972; *and references therein*). Berger (1971a) recorded the orientation and offsets within a whole set of felsite dykes to calculate the orientation of regional stresses operating on the pluton at this time.

In relation to the microgranite dykes, the majority of the felsite dykes were emplaced after, although there is believed to be some overlap, (Pitcher & Berger 1972). The felsites were definitely emplaced after the formation of the foliation within the host granite, with the present authors view being the felsite dykes were accommodating deformation when the main body of the granite was rigid.

❸ ***Pegmatites and Aplites***:- the most common minor intrusive phase within the Main Donegal Granite, notably occurring around the marginal parts of the pluton, e.g. Dunlewy-Brockagh-Croaghleconnel-Gallwolie Hill-Derkbeg-Straboy. The pegmatites have a protracted history of intrusion, some of which predate the earlier microgranites, whilst the later ones clearly post-date the felsites (Pitcher & Berger 1972). The aplites, which usually form regular bodies, (and sometimes composite dykes with pegmatites) all appear to post-date the later microgranites.

The earliest pegmatites tend to from irregular bodies which tend to grade into a host of coarser pegmatitic-type granite, which Berger (1971) believed to be the end-product of the potassic feldspathisation, responsible for producing the light banded granite due to the presence of large replacive microclines. Often these coarse pegmatites have a strong foliation marked by the alignment of quartz and biotite, although the coarse-grained nature of these rocks gives an initial undeformed appearance. There are early regular intrusions of pegmatite which are either folded or boudinaged depending on the orientation to the finite strain ellipsoid. The later dykes of pegmatite and aplite all occur as regular bodies which have not been folded or boudinaged probably because of their lack of competency contrast with the host. A mineral alignment may however be developed and is usually parallel to the dyke walls, although sometimes they may possess oblique foliations as already described. The pegmatites are composed of perthitic microcline showing a gradation into antiperthitic plagioclase. Other constituents are biotite, quartz and muscovite and occasionally garnet (Pitcher & Berger 1972).

❹ Late veins:- these minor veins or lenses, composed of quartz and epidote, all post-date the other intrusive phases listed above. The veins occupy the major joint directions within the pluton and are generally undeformed, though in the marginal areas these veins occasionally occur in a deformed state within small-scale, late stage shear zones (Pitcher & Berger 1972).

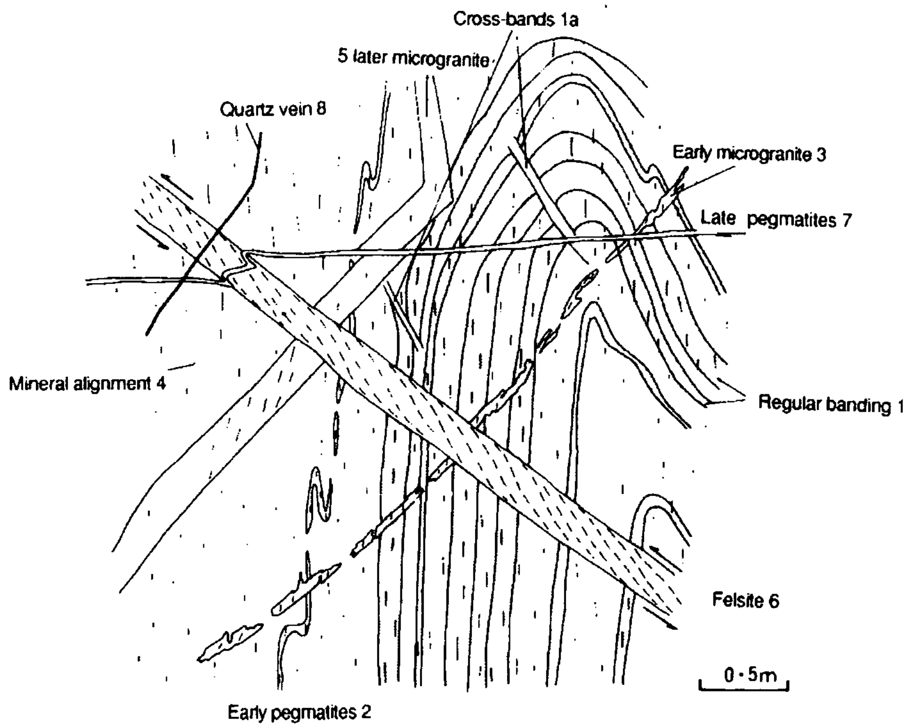


Figure 3:11:- Schematic diagram illustrating the chronology of the banding, mineral alignment and the suite of minor intrusives (*after* Pitcher & Berger 1972).

Pitcher & Berger (1972) proposed that the Main Donegal Granite and its aureole was deformed by irrotational pure shear created by a NW-SE directed regional $\delta 1$ superimposed on the pluton, possibly during, but definitely after its emplacement. The earliest event was the formation of the regular banding, generally aligned parallel to the long-axis of the pluton. Despite Bergers work, Pitcher & Berger (1972) were still uncertain about its exact origin, i.e. was it produced by i) *"in situ deformation during a highly mobile stage of the deformation"*, or ii) *"during emplacement as a result, for example, of differences in the rate of flow into place of a crystal mush"*. Coarse, diffuse, early pegmatitic patches, often folded and boudinaged, succeeded the banding. Cutting these and the earlier bands is a pervasive NE-SW trending mineral alignment, for which the maximum principle stress, σ_1 would have been approximately oriented sub-horizontal and normal to this foliation. The strong, sub-horizontal linear component, (X) indicates the direction of maximum extension, plunging NE at the north-eastern part of the pluton and SW at the south-western end. The next phase was the intrusion of a suite of microgranites which were subsequently re-mobilised by the host after emplacement. Pitcher & Berger (1972) believed that at this stage the pluton was still deforming relatively homogeneously, although by the time of the intrusion of the later microgranite dykes, (as well as some pegmatites and earlier members of the felsites) the majority of the pluton had become rigid, with deformation concentrated into localised zones of movement, which the dykes often occupied, as indicated by the presence of oblique foliations within these dykes. From measurements of orientation and sense of displacement of numerous felsite dykes Berger (1971a) interpreted the stress regime to be the result of σ_1 being oriented sub-horizontally in a NNW-SSE direction. As described, the later microgranites did not appear to be deformed within the host granite, but where these same dykes cross-cut certain enclaves, such as limestone or pelite, they were intensely deformed. This implies the deformation was still occurring within the granite, but because of the similar competency between the microgranite and the host granite the dykes were not deformed in this way. The orientation of the veins and dykes within these enclaves defined an oblate strain ellipsoid, where $1>k\geq0$. The average X/Z values of inclusions was 3 (using igneous inclusions as strain markers).

The later phases of minor intrusions are relatively undeformed in the centre of the pluton but towards the margin they are often folded and boudinaged, implying deformation continued for longer in the marginal zones of the pluton than in the centre (Pitcher & Berger 1972). Towards the margins of the pluton there is an increased component of prolate strain, with LS fabrics developing, where X (horizontal) and Y are parallel to the contact of the pluton with Z horizontal and perpendicular. The later development of the cross cleavages and the sub-horizontal striae on these surfaces was believed to attest to continuing compression from the NNW-SSE quadrants (Pitcher & Berger 1972).

This new model favoured in situ deformation of a cooling pluton rather than the forceful model proposed by Pitcher & Read (1959). Pitcher & Berger (1972) still believed the pluton to be composed of a multitude of sheets, but preferred vertical wedging, rather than horizontal migration of magma. The raft-trains represent a true ghost-stratigraphy, and were not conveyed from the NE part of the pluton by a south-westerly flowing magma. The splaying out of the raft-trains from the NE to the SW and the intrusion lineation was the strongest evidence for lateral migration, according to the Pitcher & Read model. This latter feature could be clearly explained in the new model by superimposed deformation, whilst the former phenomena (raft-splaying) was produced by the intrusion of the NE prolongation of the Thorr pluton deflecting its own country-rocks prior to the emplacement of the Main Donegal Granite (Berger 1967).

The origin of this later regional, though somewhat localised deformation event associated with the Main Donegal Granite was still not fully understood by Pitcher & Berger and nor was it clear why the deformation was localised around one of the younger members of the batholith.

The next investigation of the Main Donegal Granite was by Chenevix-Trench (1975: *unpublished thesis, Newcastle-upon-Tyne*) who had mainly concentrated on the north-western aureole of the pluton, with only limited mapping of the pluton itself. In the aureole Chenevix-Trench (1975) found evidence for simple shear due to the progressive change in the orientation of the earlier regional lineation and also the progressive rotation of the S_6 (Chenevix-Trench 1975; Hutton 1982) foliation towards the granite. The sense of rotation implied that sinistral shear had occurred within the aureole and granite. The finite strain ellipsoid implied shearing in a sub-vertical plane in a NE-SW direction. The XY plane of the strain ellipsoid lies approximately parallel to this direction which led Chenevix-Trench (1975) to suggest high shear strains are associated with the granite (i.e. during simple shear the XY (cleavage) plane will initiate at 45° to the shear direction but during progressive simple shear the XY plane rotates into near parallelism with the shear direction (Ramsay & Graham 1970)).

Within the pluton itself, Chenevix-Trench (1975) extended Pitcher & Berger's comment that the pluton was originally wider and less elongate, suggesting that the raft-trains were probably still continuous screens when the sheets were being emplaced. The extension of the pluton in a NE-SW direction has in affect caused large-scale chocolate-tablet boudinage, (i.e. Z>X=Y) of these screens breaking them up into smaller fragments. Furthermore the en echelon nature of the raft-trains was difficult to incorporate into the Pitcher & Berger model of pure shear, though the new model of sinistral shear would cause the boundaries of individual rafts to rotate into parallelism with the shear plane (Chenevix-Trench 1975). The same author believed the shear strains were highest along the NW margin of the granite due to almost all of the structures lying parallel to the contact. However, on the SE margin of the pluton the raft-trains, (the Glenleheen and Derryloaghan zones) and the foliation are more oblique to the contact, possibly implying lower values of shear strain.

Chenevix-Trench (1975) envisaged that the Main Donegal pluton was rotated anti-clockwise during simple shear around a "hinge line" occupying the present position of the strike swing (i.e. Crockator to Lettermacaward). West of this apparent hinge line the deformation is limited . The change of strike of the countryrocks from

NE-SW across the strike-swing to a NW-SE in western Donegal was thought to have been responsible for reducing shear strains in this area. Therefore the Trawenagh Bay Granite could be interpreted as an undeformed remnant of the Main Donegal Granite, with it protected from the sinistral deformation by the strike-swing and the presence of the Ardara and Rosses plutons to the south and north of it respectively (Chenevix-Trench 1975). The lack of displacement along the Trawenagh Bay-Main Donegal transitional boundary could be attributed to "hinging". If so the displacements should increase towards the NE end of the pluton.

The model of Chenevix-Trench (1975) in summary, relates the deformation within the pluton and the aureole to sinistral shear superimposed onto an essentially passively emplaced Main Donegal Granite. Chenevix-Trench believed it was passive emplacement because it was "difficult to imagine forceful intrusion resulting in intense simple shear without invoking lateral flow of magma". The orientation of the XY plane of the strain ellipsoid, almost parallel to the shear direction, was interpreted as very high shear strains within the aureole and marginal areas of the pluton, (Chenevix-Trench 1975).

3:4:4 The model of Hutton (1982)

The most recent interpretation of the Main Donegal Granite, was by Hutton (1982) who suggested the Main Donegal Granite was emplaced into an active sinistral shear zone, i.e. a true syn-tectonic granite. Strain analysis on the granite and its aureole by Hutton puts more emphasis on tectonic space creation, by which the pluton was accommodated within the shear zone rather than forcefully intruded.

The previous workers on the granite; Pitcher & Read (1959); Pitcher & Berger (1972); Chenevix-Trench (1975) believed the deformation to be localised around the granite. However, Hutton (1982) found the shear zone to extend from Fanad, where it terminated against late faults, through NW Donegal and on into southern Donegal where, just NE of Killybegs, the shear zone is overlain by Carboniferous rocks. The overall length of the shear zone is therefore in excess of 80 Km.

Along the Main Donegal Granite Shear Zone there is a displacement gradient, where the maximum displacements of 20 km are recorded in the NE end of the pluton, with this value decreasing towards the SE. The gradient was believed to have been produced by the presence of unconsolidated granite within the shear zone, locally creating high strains and subsequent instability, causing one wall to bend, with an axial crack developing. This would produce a low pressure zone within the shear zone, effectively allowing granitic magma to be "drawn in" (see figure 3:12).

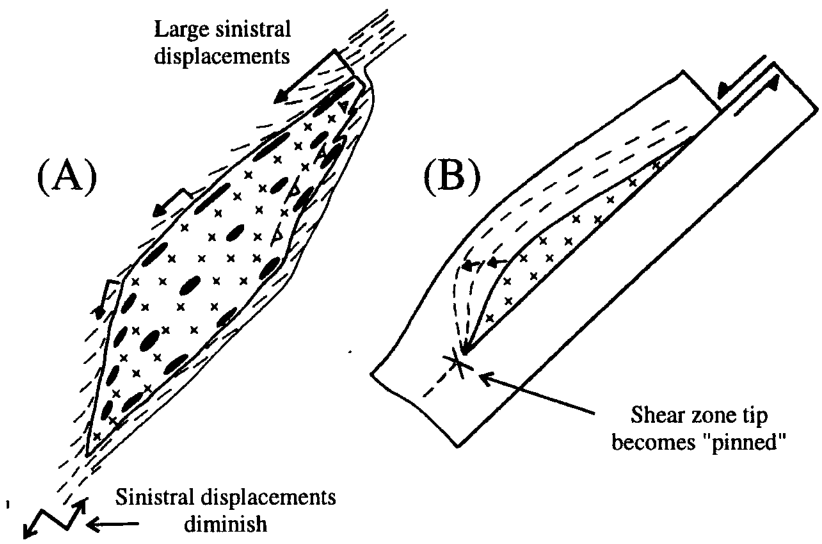


Figure 3:12:- a) Strain gradients along the length of the Main Donegal Granite.
b) Cavity created by displacement gradient along the shear zone which causes shear zone to split axially allowing emplacement of granite (*a & b* from Hutton 1988a).

The evidence suggesting syn-tectonic intrusion is as follows:-

a) the foliation within the granite is identical in orientation and morphology to the S_6 of the country rocks, (Hutton 1982; 1983), and at the NE and SW ends of the pluton the foliation is continuous across the contact.

b) F_6 folds in the aureole are cross-cut by coaxially folded pegmatites and quartz veins, with the folds and intrusions sharing the same axial planar foliation. This demonstrates at least some intrusions (if not all) are synchronous with the deformation.

c) Pitcher & Berger (1972) believed the main high grade minerals within the aureole predate the deformation, whilst Hutton (1982) found evidence that some of these minerals preserved S_6 crenulations but were overprinted and transposed also by S_6. At Lackagh Bridge there are xenoliths of mullioned quartzite within the granite. Both Pitcher & Berger (1972) and Hutton (1977) believed the mullions are related to the granite deformation.

d) In a true syn-tectonic intrusion one will expect to see a drop in strain recorded in the granite when crossing the contact. This is because the xenoliths within the granite will only start to deform when the ductility contrast between the host granite and the xenoliths has greatly diminished. Therefore the xenoliths will only record the later increments of strain, whilst the countryrocks which have remained "plastic" throughout this period will record almost all of the strain. In the Main Donegal Granite, Hutton (1982) reported such a drop in strain across the contact.

e) Plagioclase fabrics, (PFC) in the dark regular bands at the core of the Doochary Synform are parallel to the margins of the bands, whilst the foliation is cross-cutting at high angles to the bands. This shows that the areas of banding were areas of syn-plutonic deformation above the rheological critical melt percentage (RCMP), (Arzi 1978).

Within the aureole Hutton (1982) found considerable evidence for sinistral shear, (as did Chenevix-Trench), from the cleavage swing and deflection of earlier structures as one approaches the pluton. Within the pluton though, the lack of swing within the foliation was difficult to relate to sinistral shear (Hutton 1982). This author studied dyke and vein data in an attempt to distinguish between rotational and irrotational strain. The distinction relies on the principle that during progressive deformation there will be an overlap between the extension and shortening fields of the finite strain ellipsoid. With irrotational deformation (e.g. pure shear) the overlap will be symmetrical, whilst with rotational strains (e.g. simple shear), the overlap will be asymmetric, with this asymmetry indicating the sense of shear. At 8 of the 11 locations where Hutton (1982) measured the veins, sinistral rotation could be demonstrated with $k \approx 1$. The majority of the xenoliths within the granite are oblate in shape with $k \approx 0$, although towards the NE end of the pluton, the shapes are less oblate ($0<k<1$) and X is sub-horizontal. Hutton (1982) suggested that the pluton had its own additional component of strain, which was almost absent in the aureole (see figure 3:13).

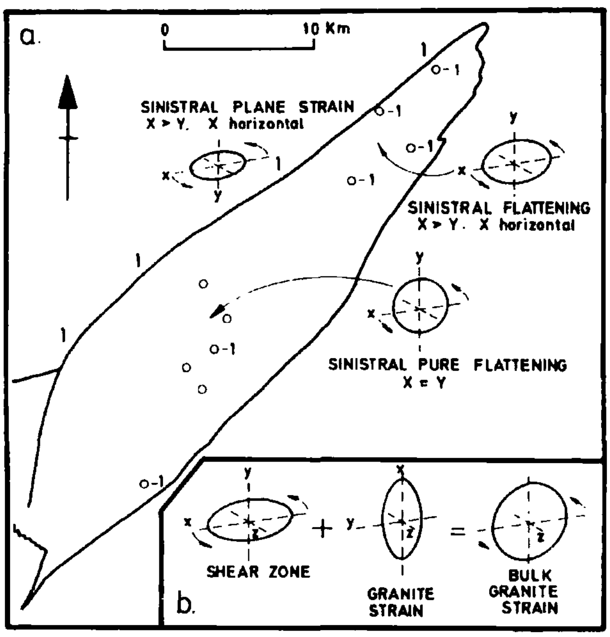


Figure 3:13
3-D strains within the Main Donegal Granite
a) k-values (K= 1 plane strain (LS); k= 0 pure flattening (S); k= 0-1 (S-LS) in granite and aureole.
b) Factorisation of bulk granite strain.
(*after* Hutton 1982)

To produce a bulk flattening within the pluton, there must be another irrotational $k \approx 1$ ellipsoid, with X vertical, coaxially superimposed on the $k \approx 1$ sinistral shear component associated with shear zone, with X generally sub-horizontal. In the margins and the NE end of the pluton the shear zone component has overcome the granite component, accounting for the strongly developed sub-horizontal lineation. Elsewhere in the pluton the two strain components have combined to produce a bulk flattening effect. The superposition of the two strains accounts for the consistent NE-SW strike of the foliation within the pluton (Hutton 1982). The double boudinage produced by the bulk flattening only extends for a few tens of metres into the aureole. This observation is why Hutton (1982) proposed that the pluton was passively emplaced because if the intrusion was forceful then the zone of bulk flattening within the aureole would be much wider. The vertical X direction within the granite strain component may be the result of buoyancy within the granite or the confining pressure of the steep pluton walls, causing vertical extension, (i.e. squeezing) of the granite. The presence of a shear zone culmination within both the granite and the aureole may be an effect of the shear zone having been vertically inflated by the granite. This culmination is indicated by a change in the plunge of the marginal lineation. Along strike of the culmination, to the SE of the pluton the Knockateen Slide is exposed at its highest topographical level in Donegal, whilst to the NW of the pluton the lowest structural levels of the D_2 tectonic slides are exposed along strike of this feature, (Hutton 1982). Assuming the plunge of the lineation is parallel to the roof then the central areas of the pluton are the most deeply exposed with ~6 km of granite having been eroded, (Pitcher & Read 1959; Hutton 1982). The heavily permeated nature of the pelite rafts in the central areas agree with this proposal.

The high, negative ε_{Nd} values for the Main Donegal Granite and the Trawenagh Bay Granite implies a contribution may have been from old radiogenic basement (Dempsey *et al.* 1990). This may be the Proterozoic gneisses of the Rhinns Complex which outcrop on Inishtrahull, a small island off the north coast of Malin Head, (Dickin & Bowes 1991).

Within the Main Donegal Granite there is uranium mineralisation, notably found in the south-western part of the pluton around Glenleheen and Cloghercor, with more sporadic occurrences within microgranites and pegmatites in the sheeted contact zones. Two types of uranium deposits are found; i) pegmatite hosted and, ii) within mineralised veins. Within the pegmatites there is low-thorian uraninite, with biotite being the main mineralogical control on its distribution (O'Connor *et al.* 1984). Furthermore where the pegmatite is close to raft-trains (the Glenleheen Zone 5) uraninite tends to be present. Textural evidence consistently attests to late magmatic formation of the uranium. The age of the uranium is 407 ±4 Ma, obtained from $^{207}Pb/^{206}Pb$ dating of uraninite, and agrees with the Rb-Sr whole-rock age obtained by O'Connor *et al.* (1982).

The second type of uranium mineralisation occurs as pitchblende bearing veins which appear to have been formed under low temperature hydrothermal conditions and is dated at 295 ±4 Ma ($^{207}Pb/^{206}Pb$ dating), possibly related to an Hercynian event, (O'Connor *et al.* 1984). The low thorian uraninite within the pegmatites may have been the source of the mobile uranium, where under suitable conditions it was transported and deposited as pitchblende veins. The occurrence of both types of mineralisation in outcrop strongly supports this. The main zone of mineralisation extends from Croaghleheen, striking parallel to the raft-trains, nearly as far Lough Muck, on the south-eastern part of the contact. The nature of the uranium mineralisation and the reason for its location in this part of the Main Donegal Granite is still not fully understood, although O'Connor *et al.* (1984) suggested the linearity of the radiometric zone may coincide with major annealed shear zones which were active during emplacement. The later pitchblende veins often trend NNW and probably exploited the joint system within the granite. The Main Donegal Granite and the other members of the Donegal Batholith are generally not associated with metalliferous ores, unlike the Hercynian granites of SW England. Whole-rock levels of uranium are usually <10 ppm, whilst thorium levels are between 10 and 20 ppm, (O'Connor *et al.* 1984; and personal data of the present author).

3:5 The aureole of the Main Donegal Granite

The aureole of the Main Donegal Granite consists of rocks which are more typical of regional metamorphism, tending to be schistose in character and composed of amphibolite grade assemblages, i.e. sillimanite, kyanite and staurolite, in contrast to the mid-upper greenschist facies regional assemblages. The structural effects can be recorded up to 5 km away from the contact to the NW of the pluton, whilst on the SE side it is narrower and bounded further to the SE by the Knockateen Slide, which appears to have acted as a decollement, recording relatively late, brittle, oblique, sinistral movement in a NNE-SSW direction, (Hutton 1982). Mineralogical changes are recorded 2-3 km from the margins of the granite.

3:5:1 Structures within the aureole

Pitcher & Read (1960) reported the development of a new crenulation cleavage which intensified towards the pluton, with an accompanied development and tightening of "new" folds. Close to the contact this crenulation cleavage becomes the dominant penetrative schistosity often exploiting earlier structural trends.

Pitcher & Berger (1972) revealed the structural story to be more complicated than Pitcher & Read (1960) had originally anticipated, implying that the folds in the aureole were of little significance to the intrusion of the pluton. The former authors believed that the increased ductility of the envelope due to rising temperatures associated with the emplacement of the Main Granite caused earlier structures to be coaxially folded by relatively minor granite related folds. The deformation was divided into three events; DMG_1, DMG_2 and DMG_3 by Pitcher & Berger (1972) and the latter two events produced the main structures now visible within the aureole. Figure 2:11 (chapter 2) correlates these phases of deformation with the more recent structural interpretation of the aureole deformation performed by Hutton (1977; 1981; 1982; 1983). Hutton (1977) studied the aureole in the Lackagh Bridge area at the NE end of the pluton and showed that there is significant folding coeval with granite emplacement. The following structures within the aureole are taken from his work (Hutton 1977). Deformation associated with the Main Donegal granite is D_4 in this area, although is regionally D_6, (Hutton 1981; 1982). In this area minor second order F_4 monoforms, (with short limbs ~200-300m long) change vergence indicating the presence of a larger F_4 structure which Hutton termed the Lackagh Synform (see figure 3:14).

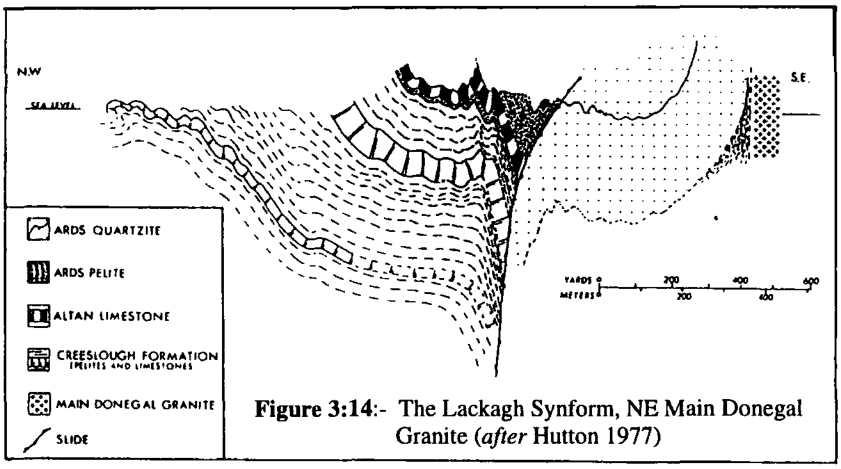


Figure 3:14:- The Lackagh Synform, NE Main Donegal Granite (*after* Hutton 1977)

Within the core of this synform there is a tectonic slide, along the contact between Ards Pelite and Ards Quartzite. Evidence for this is seen in the local vergence of minor structures close to this junction, a feature incompatible with a stratigraphic relationship. Furthermore the S_4 foliation which is usually oblique to bedding is parallel near this junction, implying high strains along this contact. Later work by Hutton & Alsop (1995) also revealed similar age slides in the Dunlewy area which are generally parallel to the pluton, all with sub-horizontal oblique stretching lineations and oblique sinistral shear sense.

Previous workers believed the NE aureole to be on the inverted limb of the Errigal Syncline, although detailed study by Hutton (1977) revealed a change in vergence of F_2 structures between the NW end of the Lackagh River and Ards Priory, hence locating the axis of the Aghla Anticline. This places most of the inner aureole on the normal, right way-up limb of the D_2 Aghla Anticline, a relationship also suggested by the stratigraphy (presence of the Altan Limestone above the Creeslough Fm.).

The "Creeslough Downwarp" of Pitcher & Berger was believed to be a pre-granite broad monoform with no associated minor structures which caused bedding, S_2 to become more steeply inclined according to these authors. Hutton (1977) related this feature to F_4 folding coeval with granite emplacement. This author believed the folds were more related to the sinistral deformation of originally gently inclined bedding and S_2 rather than the forceful emplacement of the pluton. Minor buckles would become progressively rotated anti-clockwise into parallelism with the contact as shear strains increase. Progressive rotation will result in these folds getting sheared and stretched parallel to their axes as they pass through domains of extension and compression of the strain ellipsoid causing their wavelength and amplitude to vary. The F_4 folds are broadly analogous to the DMG_2 phase of Pitcher & Berger (1972). The DMG_1 phase of Pitcher & Berger (1972) essentially relates to the coaxial and coplanar flattening of earlier structures, as indicated by porphyroblast growth. Also the mullions were believed to have formed before or during this early phase of deformation since mullioned quartzite xenoliths within the granite are sometimes overprinted by a mineral lineation parallel to the similar structure within the pluton itself. Hutton (1981) suggested that this early DMG_1 phase belonged to D_4 and stated "*the major difficulties that Berger (1980) sees in correlating the main aureole deformation with one single strain event can be explained in the context of a shear zone*". The incorporation of xenoliths during progressive simple shear would result in shear strains accumulated before incorporation being overprinted by later increments of strain (Hutton 1981). The orientation of later structures relative to early ones will be governed by the shape of the xenolith as one would expect some rotation of the xenoliths by primary flow within the granite before the viscosity contrast diminished enough for the xenoliths to start recording the strain by ductile shape changes.

D_5 (Hutton 1977) produces very minor structures which have little affect on the overall structure. D_6 produces the extensional cleavages within both the granite and aureole (the cross-cleavages of Pitcher & Berger 1972). The dominant orientation is NNE-SSW associated with sinistral offests with a less common conjugate pair (dextral offsets) approximately oriented E-W. The dominance of the former cleavage also suggests sinistral shear in the aureole and margin of the pluton. D_7 structures are sporadically developed contractional kink bands, mainly in the aureole and occasionally in the granite. Pitcher & Berger (1972) grouped D_5, D_6 and D_7 structures as DMG_3. They also included small localised shear zones within the aureole, where strong S-C fabrics are developed. These belts are usually 100-200 metres wide and are found on both sides of the pluton, e.g. the Binnadoo Shear Belt around Lough Greenan. These same type of structures may be analogous to the "Medial shear zones" of Hutton & Alsop (1995) which they report in the Dunlewy area. Sense of shear is sinistral along these zones.

In the Gweebarra Bay area the cross-cutting relationships and metamorphic textures consistently show that the Main Donegal Granite and its deformation are younger than the Ardara Pluton. Within all rock types there is a very strong SW plunging mineral stretching lineation developed in all rock types including the appinites, quartzites and the "stalk" of the Ardara Pluton (Iyengar *et al.* 1954). Outside the shear zone on Crockard Hill, X/Z ratios increase from 5.0 to 7.6 inside the shear zone on Meenalargan Hill (Meneilly 1982). In this area deformation associated with the Main Donegal Granite was termed D_5 by Meneilly (1982) due to the prominent D_4 phase of deformation synchronous with the intrusion of the Ardara Pluton, (see figure 2:11). Meneilly (1982) observed the orientation of F_5 from WNW-ESE away from the granite to NE-SW at the contact. This situation can be explained by progressive simple shear as away from the granite folds will first form at 45° to the shear direction, but as shear strains increase towards to contact the folds will be rotated into near parallelism with the shear direction (Meneilly 1982).

3:5:1:2 Deformation away from the Granite

The Main Donegal Granite Shear Zone has been traced from the NE of the pluton to the west Fanad area where it is believed to terminate as a transcurrent imbricate stack, producing a compressional bulge in the form of a domal structure (White & Hutton 1985). Within this area the Knockateen Slide is affected by the shear zone with this major structure acting as a decollement to the shear zone deformation. This is evident from the intense NE-SW trending S_6 foliation and sinistral S_8 within the Creeslough Succession below the slide, whilst in the overlying Kilmacrenan Succession S_6 is only weakly developed, and S_8 absent. The only strong evidence for sinistral shear on the shear zone in West Fanad is the strong development of sinistral extensional crenulation cleavage. The major structures are a series of north-westerly verging F_6 folds separated by two transcurrent tectonic slides which can be identified by zones of intensely developed blastomylonites. The plunge of these folds is parallel to the mineral stretching lineation developed on S_6 cleavage surfaces. The bounding nature of the tectonic slides to either side of the folds is analogous to an imbricate transcurrent stack. On crossing the Knockateen Slide the deformation has been dispersed along this older structure, although some shortening has been accommodated by a broad open warp of the slide itself producing a low amplitude fold trending approximately N-S. This domal structure is termed a "compressional bulge" by White & Hutton (1985) associated with the termination of the Main Donegal Granite Shear Zone.

3:5:2 The Metamorphic History of the Main Donegal Granite Aureole

The mineralogical effects associated with the intrusion of the granite are confined to a 2-3 km zone bordering the pluton. The isochemical nature of the metadolerites shows the increase in metamorphic grade changing from fine-grained, massive (though often heavily foliated) plagioclase, hornblende and chlorite bearing rocks away from the pluton, to fresh looking medium-grained schistose amphibolites, containing hornblende, oligoclase and sphene at the contact and within rafts in the pluton (Pitcher & Berger 1972). Figure 3:15 shows the mineralogical changes within certain lithologies within the aureole of the pluton, plus rafts within the granite. The most detailed account of the metamorphic history of the aureole is contained within Pitcher & Berger (1972) who summarise earlier works of Pitcher & Read (1963) and Naggar & Atherton (1970). Structures within porphyroblasts, (i.e. inclusion trails etc.) are termed S*i* fabrics, whilst external fabrics are termed S*e*.

Within some parts of the aureole there is a distinction between an inner and outer aureole, e.g. in the Crockmore area the Binadoo Shear Belt, (DMG_3 of Pitcher & Berger 1972; D_8 of Hutton 1982) separates chlorite bearing schists in the SE from biotite schists containing andalusite, sillimanite, kyanite and staurolite to the NW. A similar situation is seen in the Fintown area where andalusite is restricted to the Glencolumbkille Pelite, whilst NW towards the contact staurolite is predominant (Pitcher & Berger 1972). In general though, andalusite is observed well outside the sheeted zone, whilst kyanite, sillimanite, (and fibrolite) are seen closer to the contact and within the rafts.

The growth of early porphyroblasts such as staurolite, kyanite, oligoclase, sillimanite, andalusite and garnet was generally believed to be an initially static event, although during the DMG_1 event (D_6 of Hutton 1982) there was period of flattening which produced a discordance between the S*i* and S*e* fabrics (Pitcher & Berger 1972). The event was also marked by recrystallisation and general matrix coarsening.

The earliest porphyroblasts are kyanite, staurolite and plagioclase, which all tend to show similar textural relationships. Often these minerals contain inclusions, S*i* of quartz, mica and graphite which are generally finer grained than their S*e* counterparts, supporting the evidence for matrix coarsening during this event. Kyanite, plagioclase and staurolite tend to overprint early crenulations which were subsequently tightened around the porphyroblasts to produce the strongly external schistosity, i.e. the initial growth was "static", (Pitcher & Berger 1972). Often the S*i* trails were sigmoidal. Pitcher & Read (1963) stated *"the restriction of a single*

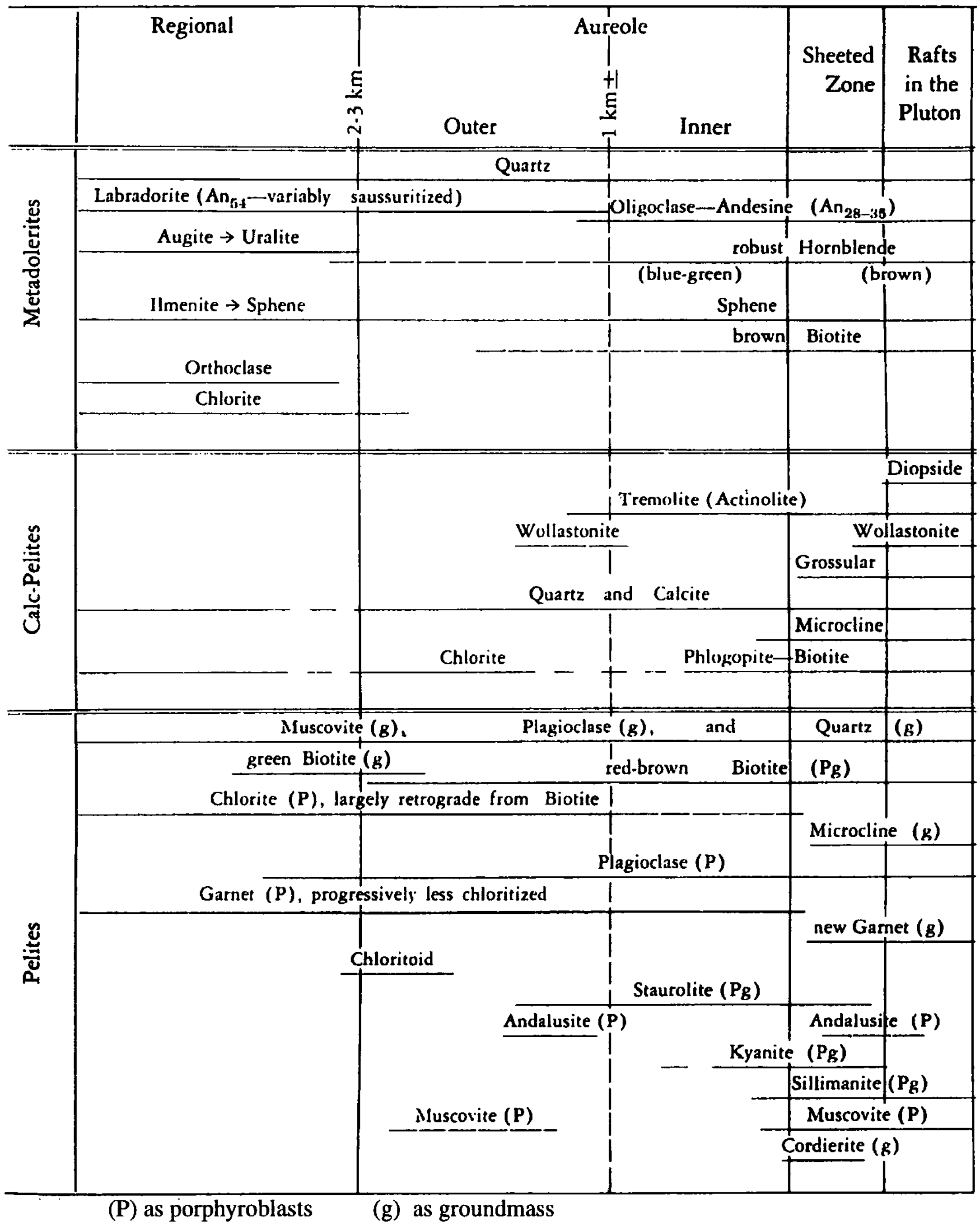


Figure 3:15:- Variations within mineralogy of some of the envelope rocks on approaching the Main Donegal Granite (*taken from* Pitcher & Berger 1972).

sigmoid to each of the porphyroblasts without regard to size strengthens our opinions that the porphyroblasts were rotated during the formation of microstructures". Pitcher & Berger (1972) argued against this point saying the lack of symmetry of the S*i* patterns around the centre of the host crystal put doubt on their syn-tectonic growth and stated the discordance between S*i* and S*e* was produced by flattening. In contrast, Hutton (1982) agreed with Pitcher & Read (1963) believing the rotation of the porphyroblasts was due to D_6 deformation with the S*i* fabrics representing earlier increments of sinistral shear, overprinted by later increments of the same D_6 event. Garnet appears to be related to two phases of growth. The earlier larger garnets form chloritised porphyroblasts and are occasionally enclosed in plagioclase and andalusite crystals and appear to be remnants of the peak regional metamorphism, (D_2, pre-D_3, Hutton 1983). Later, fresher, smaller garnets appear to be coeval with the thermal kyanite, plagioclase and staurolite

Andalusite tends to form porphyroblasts often larger than the other porphyroblasts already described and in some occasions the andalusite contains oligoclase, kyanite and staurolite as inclusions. This relationship may imply that it either formed later, or it outlasted the growth of the other minerals, (Pitcher & Berger 1972). The lack of replacement textures between andalusite and kyanite suggests they grew independently of each other with the composition of the host rocks a controlling factor (Pitcher & Read 1963; Naggar & Atherton 1970).

The distribution of sillimanite in the aureole is not fully understood as there appears to be two periods of growth. Sillimanite prisms are only found at the immediate contact or within the rafts, whilst fibrolite is much more widespread, occurring within and just outside the sheeted zone. The majority of the sillimanite and fibrolite shows a preferred orientation lying along the S_2 foliation (of Hutton 1982). Furthermore there are mats of fibrolite folded by DMG_2 (still D_6 of Hutton 1982), although occasional needles of fibrolite lie along the S_6 cleavage, implying that fibrolite and sillimanite may have preceeded D_6. Around andalusite porphyroblasts fibrolite sometimes forms fringes, implying it is later than the majority of the porphyroblasts already discussed. Pitcher & Berger (1972) noted the common occurrence of sillimanite with late muscovite.

Following the growth of these early minerals there appears to be another episode of porphyroblast growth, mainly phyllosilicates. Near the contact of the pluton muscovite is common, whilst biotite dominates the inner aureole with chlorite present within the outer aureole, (Pitcher & Berger 1972). Often biotite is replaced to varying degrees by chlorite, although this may be interpreted in two ways:- i) the retrogressive replacement of biotite by chlorite, or, ii) prograde reaction where chlorite is being replaced by biotite. The later phyllosilicate porphyroblasts are separated from the earlier ones by the flattening event and matrix coarsening, as indicated by overprinting textural relationships. In regards to the deformation within the aureole this later porphyroblast event predates the bulk of the DMG_2 event, (D_6 of Hutton 1982), though there is the possibility of some overlap between the two.

3:5:2:1 The late retrogression

Within the above mentioned assemblages there is a general retrogression associated with the later part of DMG_2 and DMG_3, (D_{7-9} of Hutton 1982) visible throughout the aureole but more prevalent within the outer aureole, (Pitcher & Berger 1972). In these outer areas biotite is often chloritised, andalusite is replaced by fibrous white mica, whilst kyanite and staurolite are partially altered to muscovite; plagioclase is often seriticised, and the garnets are altered to chlorite, although the smaller garnets mentioned earlier are generally less affected, (Pitcher & Berger 1972). The presence of chlorite along the DMG_3 cross cleavages, (D_8 of Hutton 1982) correlates the retrogression event with the later stages of deformation. The retrogression may well be associated with the cooling down of the pluton and the release of late volatiles from the granite into the aureole rocks. Figure 3:16 shows a correlation of metamorphic mineral growth and tectonic events related to the emplacement of the Main Donegal Granite (Pitcher & Berger 1972).

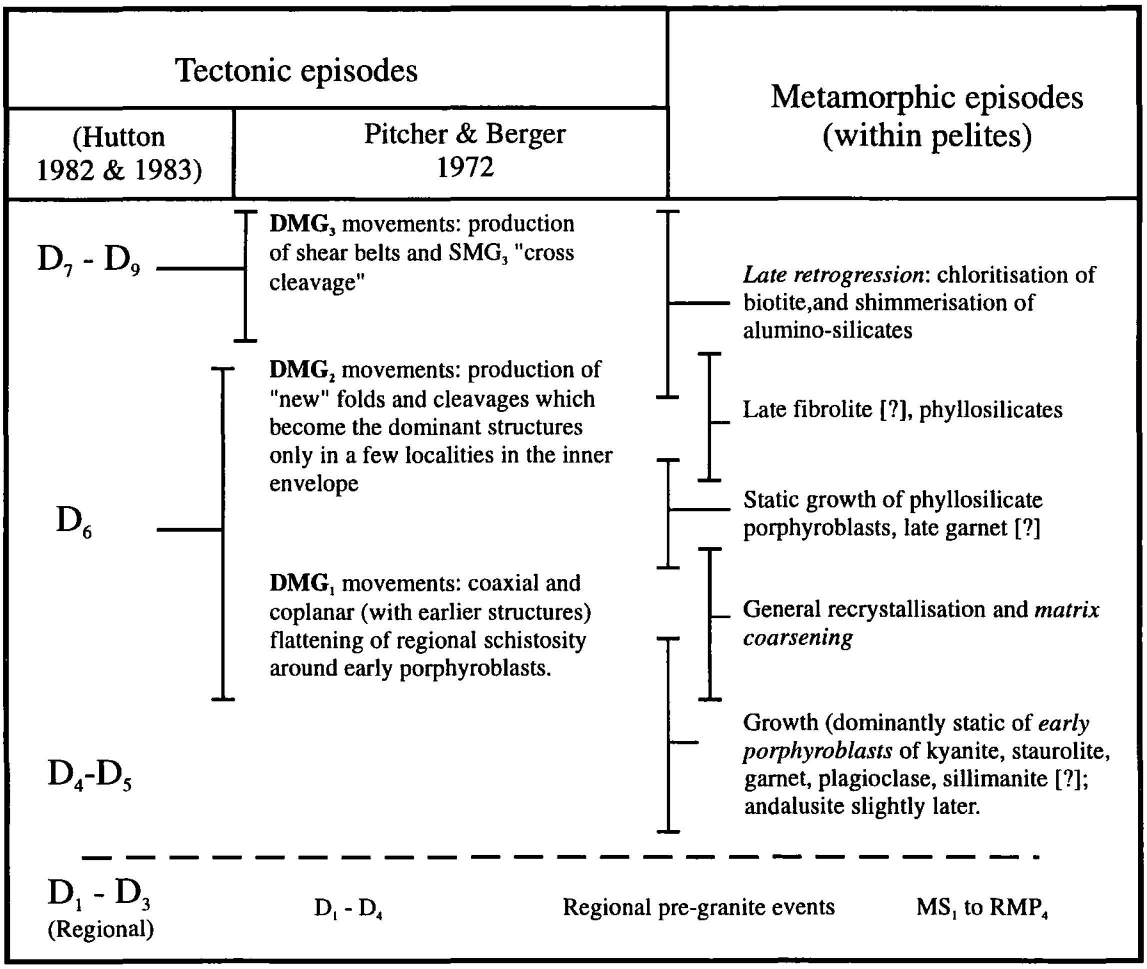


Figure 3:16:- Correlation of metamorphic and structural events within the envelope of the Main Donegal Granite (*modified from* Pitcher & Berger 1972).

3:6 The Trawenagh Bay Granite

The Trawenagh Bay Granite (pluton) forms a rectangular lobe to the west of the Main Donegal Granite, occupying an area of approximately 50 km^2 (see figure 3:17). Cross-cutting relationships demonstrate that the Trawenagh Bay body, (dated at 405 ±3, by Rb-Sr whole-rock isochron (Halliday *et al.* 1980)) is younger than the Thorr Pluton, whilst along its northern contact it truncates the G1 member of the Rosses Pluton.

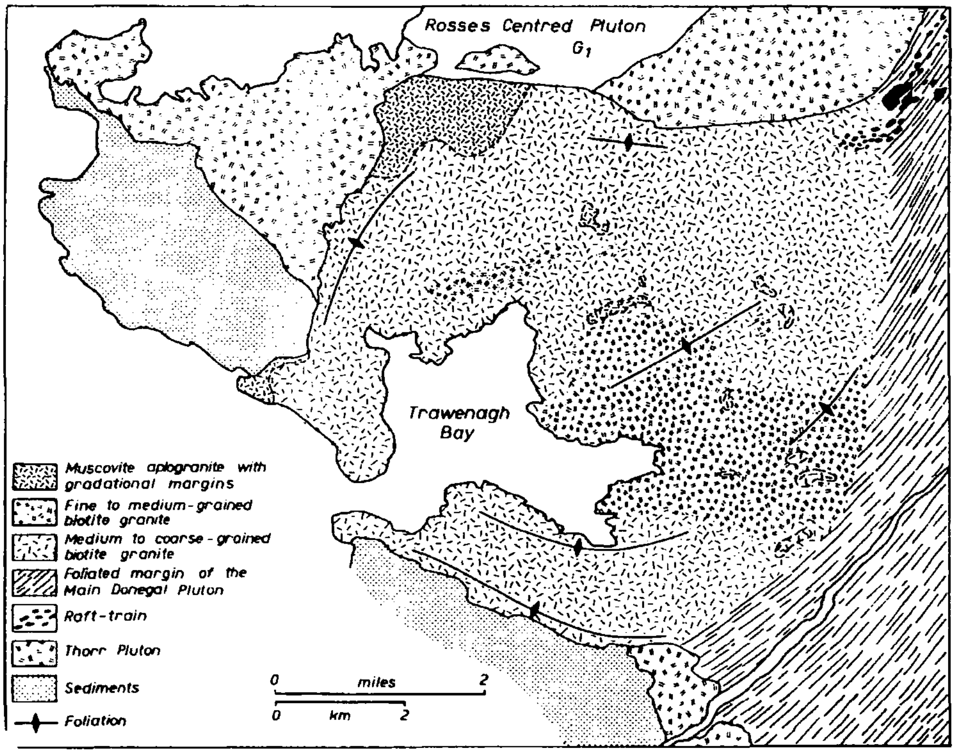


Figure 3:17:- The Trawenagh Bay Pluton (*after* Pitcher & Berger 1972)

The junction between these granites is sharp, although difficult to see due to their similarity; the strongest line of evidence for two separate granites being the truncation of members of the Rosses porphyry dykes by the Trawenagh Bay Granite. The above relationships are clear-cut in comparison to the age relationship between the Main Donegal Granite and the Trawenagh Bay Granite. Pitcher & Read (1959) believed the latter pluton was younger due to granite sheets resembling those of the former being truncated by the Trawenagh Bay Granite in the Meenatotan area. Otherwise the contact is generally transitional in character with the characteristic features of the Main Donegal Granite, i.e. banding, mineral alignment and preferred orientation of xenoliths, dying out westwards over a distance of 0.5-1.0 km into the almost structureless Trawenagh Bay Granite. Pitcher & Read (1959) believed that the large volumes of pegmatite (most typical of the marginal areas of the granite) which form a belt extending from Brockagh southwards through Croaghleconnel and Galwollie Hill, were part of the roof extending down into the pluton but dying out at depth. It was this pegmatite fringe that the above authors believed had shielded the Trawenagh Bay Granite from the bulk of deformation associated with the Main Donegal Granite.

The majority of the Trawenagh Bay body is composed of a medium to coarse equigranular biotite granite (often muscovite bearing). In the central areas of the pluton there is a finer grained mass of granite often having diffuse contacts with the host granite, (Pitcher & Berger 1972). The most striking variant is a garnetiferous, muscovite-rich aplitic granite forming sizeable areas in the north-western and western regions, as well as smaller sporadic patches in the central areas of the granite.

The biotite granites are modally very similar despite the varying grain sizes exhibited and consist of plagioclase (An_{10-15}) (average proportions being 31%); microcline (31%), quartz (27%), biotite (4.5%), though the latter is often altered to chlorite or muscovite. Epidote forms the main accessory (Pitcher & Berger 1972). These normal biotite granites grade into aplogranitic variants, a transition marked by the disappearance of biotite, accompanied with increasing proportions of muscovite and then garnet (almandine-spessarite). Pitcher & Berger (1972) believed this garnetiferous variant was typical of marginal type granites whose crystallisation may have been influenced by the presence of volatiles, i.e. near the roof. For this reason the above authors believed that the presence of similar patches of this garnetiferous granite in the central areas meant that the granite had a relatively flat roof, quite close to the present level of exposure.

The margins of the granite are very sharp, (except for the boundary with the Main Donegal Granite) tending to be highly planar and steeply dipping. Unlike the Main Donegal Granite the Trawenagh Bay Granite does not possess a sheeted margin nor has it any minor apophyses. Structural affects related to the intrusion of the granite are absent within the adjacent envelope, with there only being very minor mineralogical changes with local recrystallisation and the random growth of muscovite porphyroblasts at the immediate contact, clearly overprinting fibrolite associated with the Thorr pluton (Pitcher & Berger 1972). Within the adjacent Thorr and Rosses granites the only effects are the slight discolouring of biotites.

Within the Trawenagh Bay Granite the foliation is generally very weak to almost absent. Moving westwards from the Main Donegal Granite the associated strong NE-SW foliation becomes much fainter swinging into a WNW-ESE trend broadly parallel to the contacts of the pluton. Although difficult to determine the fabric is dominantly planar ($0 \leq k < 1$), steeply dipping to vertical in attitude (Pitcher & Berger 1972). King (1966) studied the fabric of the pluton using the anisotropy of magnetic susceptibility and identified the existence of weak, gently SW plunging lineations, often invisible in the field.

Xenoliths of country-rocks and older plutons are generally absent from the Trawenagh Bay Granite tending to decrease in abundance in a westwards direction and where they are encountered they generally have a random orientation.

3:7 The age relationships between the Trawenagh Bay Granite and the Main Donegal Granite

This argument centres on whether the Trawenagh Bay Granite is either younger or older than the Main Donegal Granite or they are the same age but the former pluton has escaped the intense deformation which the latter suffered. The following options will be discussed using the evidence of previous researchers to argue for and against.

Pitcher & Read (1959) believed the Trawenagh Bay Granite to be the youngest. Part of their evidence was the truncation of Main Donegal Granites sheets at Meenatotan by the Trawenagh Bay body. Berger (1967) claimed this evidence was inconclusive as these sheets may belong to the Thorr or Rosses plutons. Other evidence was the prominent raft-train composed of Thorr Granodiorite and pelite at Meenderryherk (NE part of Trawenagh Bay Granite) whose xenoliths lose their strong alignment as it enters the less deformed granite; implying that the Main Donegal Granite is older (Pitcher & Read 1959). Berger (1967) claimed if this was the case then the internal structures within the unaligned xenoliths should still be preserved, i.e. the mineral alignment and deformed veins. In this area this is not the case, with the mineral alignment getting weaker and the dykes within the xenoliths become less deformed as one moves westwards from the Main Donegal Granite. Therefore the xenoliths could be interpreted as either being relicts of the older Trawenagh Bay Granite intruded by the Main Donegal Granite, although there is the possibility of this relationship being vice-versa.

The general reduction in the intensity of the mineral alignment as ones enters the Trawenagh Bay Granite could be interpreted as: i) it was too competent to transmit stress, (i.e. older), ii) it was too ductile, (above the RCMP) to transmit the necessary stresses, (i.e. younger), or, iii) the strains decreased due to it being further away from the locus of deformation, and therefore it was the same age (Berger 1967). The transitional nature of the contact between the two granites may favour the latter of these three options, i.e. where they were intruded at the same time but the deformation of the Main Donegal Granite imposed the structural features it now possesses and the Trawenagh Bay portion escaped.

Geochemical and isotope analysis tend to suggest that the Trawenagh Bay granite and the Main Donegal Granite are similar. Pitcher & Berger (1972) argued that the bulk composition of these two granites, (as well as Rosses) were very similar, and suggested they formed under lower water pressures than the other members of the Donegal Complex. An initial ε_{Nd} value of -7.5 for the Trawenagh Bay Granite and ~8.0 for the Main Donegal Granite is strong evidence to support they are possibly the same (Dempsey *et al* 1990).

This problem of the age relationships between the two granites will be addressed in the light of new work performed by the author along the transition between the two granites and by joint work with W.S. Pitcher over some key areas within the Trawenagh Bay Granite.

In summary, the Trawenagh Bay Granite is relatively homogeneous with the general absence of any accommodation structures within its aureole. This suggests the pluton was passively emplaced, although the exact mechanism of intrusion is uncertain.

4:8 Events post-dating the emplacement of the Donegal Batholith

After the emplacement of the Donegal Granites in the late-Silurian to early-Devonian, during the regional uplift of the Caledonides, an episode of wrench faulting affected much of NW Donegal. On a larger scale this faulting affected much of the Caledonides of Ireland and the British Isles. In NW Donegal there are a whole series of NE-SW trending faults of which the most important: the Leannan Fault, has already been discussed in the previous chapter. Displacements along the faults are variable, usually between 34 Km to <0.5 Km (Alsop 1992a). The larger faults are illustrated in Figure 2:6. The majority of the faults have sinistral displacements, but a few do record dextral offsets, e.g. the Gweebarra Fault, which forms a prominent topographic feature, and displaces the Main Donegal Granite by 0.5 Km.

Contemporaneous with faulting was the erosion of the Donegal Highlands, and the deposition of Devonian red beds (Pitcher & Berger 1972). In Donegal the only remnant of "Old Red Sandstone" deposition is preserved as a small outlier at Ballymastocker on the Fanad peninsula. The deposit consists of 250m of coarse clastics, composed of material mainly derived from the Dalradian (Pitcher & Berger 1972). No clasts of the Donegal Granites are found within the deposit implying:- i) the granites had not been unroofed, or ii) the area of deposition and provenance of the sediments was to far away from the outcrop of the Donegal Granites. The presence of the Leannan Fault along the southern margin of the deposit may imply the sediments were originally deposited further away from the Donegal and they are related to the faulting (e.g. as a pull-apart). Pitcher & Berger (1972) believed the granites were unroofed by the late-Devonian although absolute evidence is inconclusive due to the general absence of Devonian sediments in NW Donegal.

In southern Donegal there are marine sediments of Lower Carboniferous age associated with a marine transgression during Visean times (Pitcher & Berger 1972). The sediments are in a basin called the Donegal Syncline. In the north of the outcrop there are coarse conglomerates and gravels which show a rapid facies change into shales and limestones, (Pitcher & Berger 1972). The environment of deposition is interpreted as deltas prograding southwards into relatively shallow seas. Pitcher & Berger (1972) believed that much of NW Donegal at this time had positive topography. During the deposition of the Carboniferous sediments the strike-slip movements along the NE-SW faults had mainly ceased though some of these faults still had a component of dip (+oblique) slip, e.g. the Belshade Fault (Pitcher & Berger 1972).

Within Donegal there is little evidence of the events which took place between the Carboniferous and Tertiary, with the majority of it having been eroded. In Antrim, rocks belonging to the Permo-Trias and the Cretaceous are preserved underneath the Tertiary lavas. In Donegal there are numerous NW-SE trending olivine dolerite dykes of Tertiary age. There are two main swarms within Donegal, one of which is centred on the Main Donegal Granite whilst the other is centred on the Barnesmore Granite. Pitcher & Berger (1972) believed deep-seated fractures and joints within these plutons acted as conduits to the ascending basaltic melts. Hutton & Alsop (1996) claim the dyke-swarms are most dense where they cross the NNE-SSW lineament implying some structural control on their ascent. It is uncertain whether or not these dykes supplied magma to lava flows at higher levels because there is no evidence remaining in NW Donegal of any such flows (Pitcher & Berger 1972).

During the Quaternary the landscape of Donegal was greatly modified by the Pleistocene glaciation events. Subsequent to these glaciations large boreal forests grew over most of this area. The warming of the climate saw the gradual replacement of these forests with thick deposits of peat, which often preserve remains of the forests which once covered this area (Pitcher & Berger 1972).

The following chapters will now describe the internal variation within the Main Donegal Granite based on a combination of detailed and reconnaissance mapping.

Chapter 4

The heterogeneous granite zones in the central regions of the Main Donegal Granite

4:1:- Introduction

The aim of this, and the following chapter, is to describe the internal structure of the Main Donegal Granite based on a combination of both detailed and reconnaissance mapping in well exposed regions of the pluton. This new research has revealed the hitherto unrecognised heterogeneity within the granitoids of the pluton. At least seven major plutonic units have been identified (see figure 4:1 or Map D of this thesis). All of these units are elongated NE-SW, parallel to the long axis of the pluton and all possess a basic, lens-like geometry. These major units are commonly, but not always separated by the extensive "raft-trains" of country rock and older plutons. The larger units are characterised by general internal homogeneity and the general absence of country-rock material or older granitoid phases within them. These units characterise the NW and SE (at the north-eastern end) parts of the pluton and will be discussed in the chapter 5. In contrast, the major units in the medial regions of the pluton are more complex and are often composed of numerous granitoid facies ranging compositionally from early granodiorites and tonalites to younger porphyritic, and to a lesser extent aphyric monzogranites. This broad evolutionary range in composition-chemistry together with the field evidence implies a complex intrusion history, with these heterogeneous units representing possible sites of longer-standing intrusion. It is these heterogeneous granite "packages" within the Main Donegal Pluton that form the basis of this chapter.

Three areas have been mapped in detail in this zone:- i) Crobane Hill, ii) the Sruhanavarnis Valley, and iii) Glendowan Mountains (Croaghacullin-Moylenanav). Following their description and analysis these areas will be compared with other areas mapped along strike in a more reconnaissance fashion to establish lateral

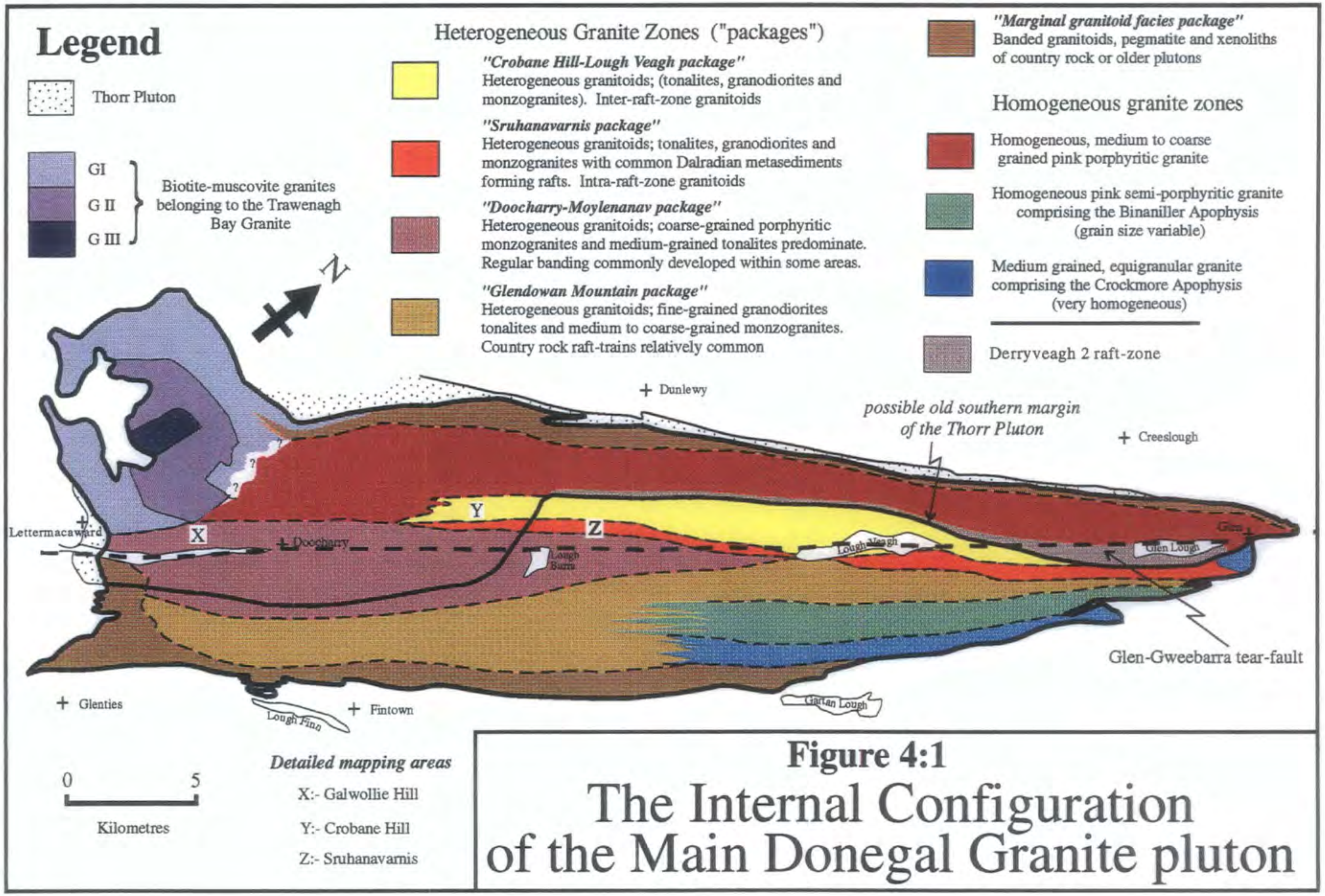


Figure 4:1

The Internal Configuration of the Main Donegal Granite pluton

continuity of the units. In the final section all the granitoids seen within the heterogeneous zones will be compared and contrasted.

Within all three of these areas detailed thin-section petrographic analysis was undertaken to support the field identification of rock types. This has been allied with geochemical analyses to identify the relationships between major element distribution and the mineral phases present. The geochemical and detailed petrographic (i.e. mineral compositions, textures etc.) aspects will be addressed in chapter 7. In order to avoid unnecessary description in this chapter the appearance of granites will be described at the field and hand specimen scale only and its typical appearance within the field. Modal proportions and compositions plus deformation structures determined from thin-section, however will be addressed in this chapter. The names of the granites in this thesis have been derived from modal analysis and using the IUGS-Streckeisen modal classification with these names being maintained, from here onwards, throughout this thesis. Where the granites are being discussed in more general terms, the less generic name of "granitoid" will be used.

4:2 Crobane Hill

Figure 4:2:- The exposure on Crobane Hill, looking south-westwards from the summit towards Gweebarra Estuary and Croaghleconnel.

Introduction

Crobane Hill lies at the south-western end of the prominent Derryveagh Mountains, approximately 3 km SW of Slieve Snaght (678m), the second highest peak in Donegal. This small hill, consisting of two well exposed glacial pavements, is most easily approached from the small hamlet of Commeen, 1.25 km to the SW, which is situated along the Doocharry-Glendowan road, (the R254). Crobane Hill lies between two major raft-zones, situated to the SE of the Derryveagh 2 raft-zone and to the NW of the Glenveagh 3A raft-zone, (see figure 4:3). The granitoids between these two zones are generally heterogeneous and range in composition from tonalites-trondhjemites, granodiorites to monzogranites.

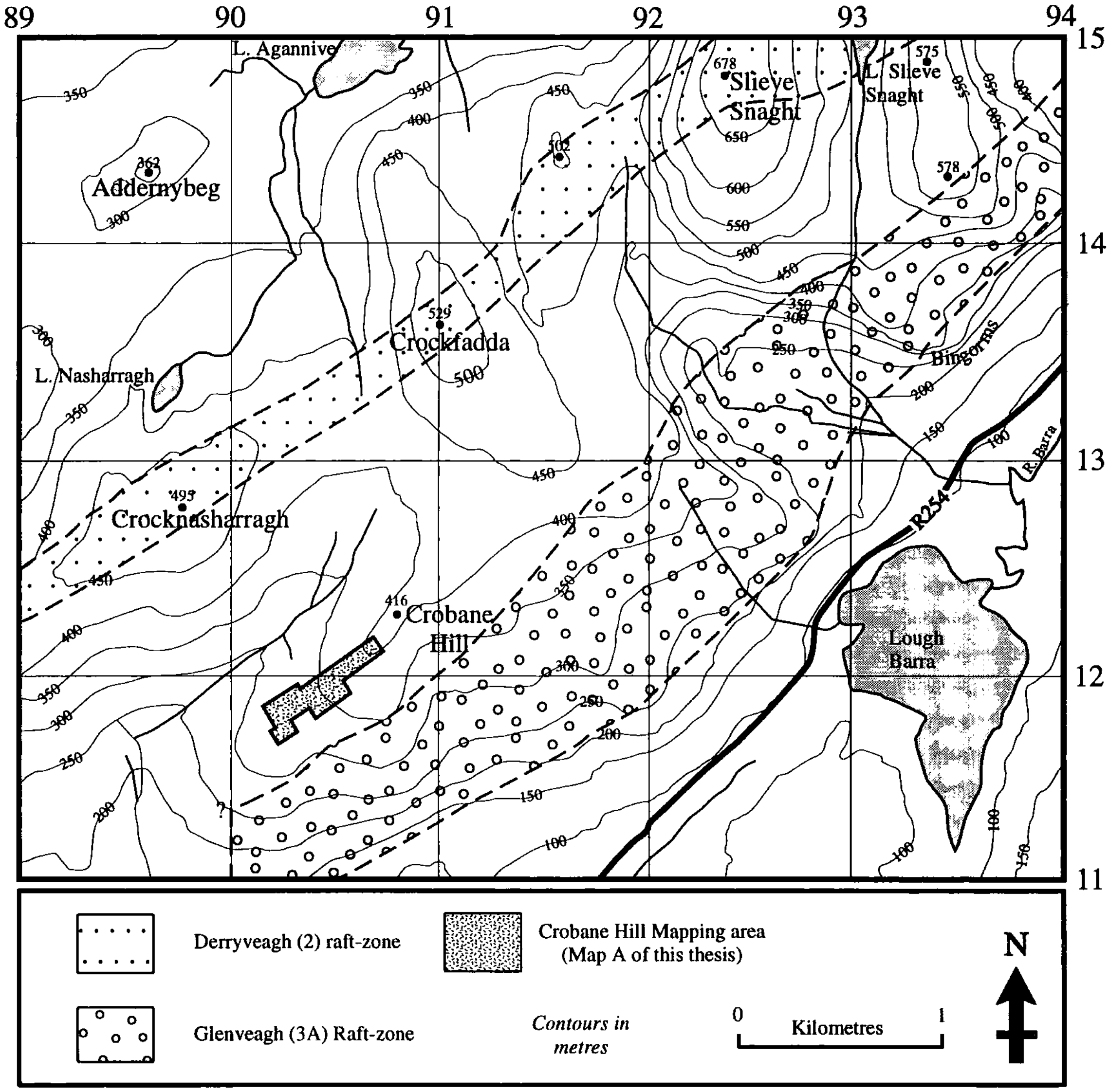


Figure 4:3:- the location of the Crobane Hill mapping area and its position relative to major raft zones.

Exposure on this hill is excellent, in terms of both degree (almost 100%) and quality, with the different granitoid types mostly identifiable from their weathered appearance. Detailed exposure mapping on a scale of 1:250 was performed in this area over a period of 3 weeks (Map A of this thesis). The map of this area covers a distance of 570 metres broadly parallel to the long axis (and foliation) of the pluton. The width of the transect is on average 180 metres and its narrower width was governed by the decrease in quality and percentage of exposure to either side of the watershed on the hill. The total area mapped was thus 54,000 m^2, an area thought sufficient in size to ascertain the age relationships between the granitoid facies present. The SW end of the transect starts at GR B 902117 continuing in a north-easterly trend towards GR B 907121. See figure 4:1 for the general location of Crobane Hill mapping area. This mapping area was marked out into squares of 30x30 m in size using a compass and tape measure. The exposures within the squares were marked out using the pacing method where the corners, (either marked by stone cairns or bamboo canes) were used as reference markers to insure a moderately high degree of accuracy. The position of the markers forms a gridline system from which all co-ordinates are derived. These co-ordinates will be quoted for areas which form the basis of detailed discussion in subsequent sections of this chapter. The gridlines are oriented 065°-245° and 155°-325° from grid north. The following sections will describe contact relationships and general features associated with emplacement observed in this mapping area.

4:2:1 Granitoid facies on Crobane Hill

On Crobane Hill there are three main granitic phases. These are cut by two phases of microgranitoid, plus minor pegmatites and aplites. Despite this area lying between two major raft-zones occasional xenoliths of quartzite, pelite and Thorr granodiorite are encountered. The relationship of the various granitoids to one another are shown in Map A at the back of this thesis. The aim of the following section is describe these granitoids in detail in regards to their appearance in outcrop and, where relevant, in thin-section. This will be followed by a discussion on important contact relationships between the different granitoid types to allow a chronology of emplacement on Crobane Hill to be constructed. The granitoid phases will be prefixed by "CBH" after the type locality and discussed in order of relative age. For most of the granitoid types modal analyses have been performed and plotted on a QAP diagram (see figure 4:4). Photographs of polished hand specimens are illustrated in Appendix A of this thesis

CBH 1

The CBH 1 granitoid plots in both the tonalite and granodiorite field on the QAP plot, although the majority of the samples are within the granodiorite field. This granodiorite tends to form lens-like pods which are often surrounded by a host of younger CBH 2 or CBH 3. The long axes of these "pods" are sub-parallel to the mineral alignment. Smaller xenoliths of this granodiorite tend to be "smeared out" in appearance to produce a streaky type of banding. The overall outcrop relationships imply that the CBH 1 granodiorite forms "rafts" within younger granitoids on Crobane Hill. The CBH 1 granodiorite is composed of white plagioclase (45%), grey quartz (28%), white microcline (14%), dark green biotite (11%), muscovite (1.5%) and epidote + others (zircon and apatite) (0.5%). In hand specimen it has an equigranular appearance, although the presence of generally larger K-feldspar megacrysts which often enclose smaller crystals of plagioclase, biotite and quartz gives this granodiorite a coarser appearance than is actually the case. CBH 1 has a dark-grey colour easily distinguishable from later granitoid phases due to its higher colour index (~ 13%) and slightly finer grain size. The grain size is generally less than 3 mm, with plagioclase crystals having an average length of 3-4 mm, with only K-feldspar larger in size.

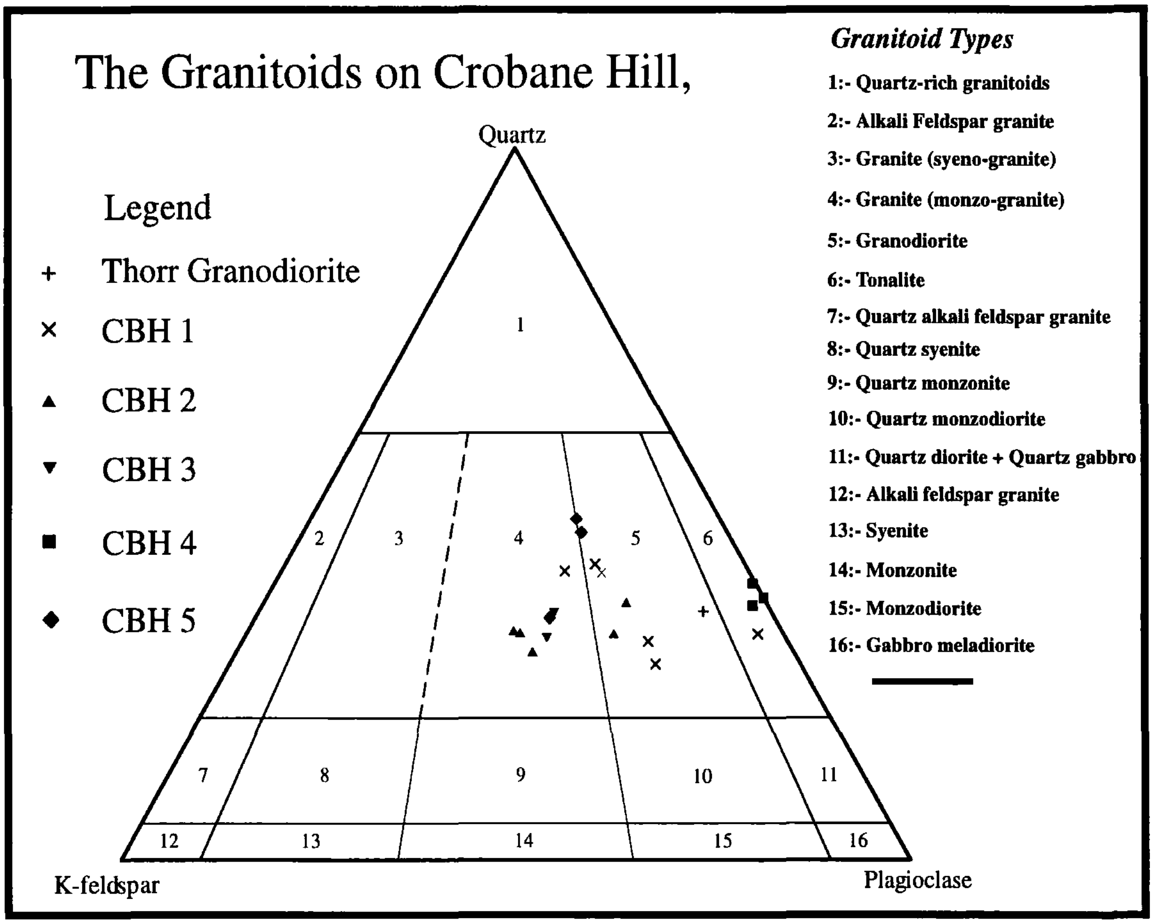


Figure 4:4:- QAP plot of the granitoids on Crobane Hill.

In certain areas the CBH 1 granodiorite does show minor variations mainly in grain size and also in the size of the K-feldspar megacrysts. The latter feature is particularly true in areas B13 and B14 where on weathered surfaces this granitoid has a similar appearance to the coarser, porphyritic CBH 2 monzogranite.

CBH 2

This porphyritic monzogranite tends to form larger, more continuous, exposures than the CBH 1 granodiorite. Within this monzogranite there are also inclusions of pelite, quartzite, Thorr Granodiorite and CBH 1, a relationship clearly observable in E7, E8 and E9. CBH 2 has an overall coarser appearance than CBH 1 mainly due to the presence of large K-feldspar megacrysts up to 15-20 mm in diameter. The mineralogy is, in order of abundance: white plagioclase (35.5%), grey quartz (30.5%), white K-feldspar (sometimes pink) (25%), dark-green biotite (7.5%), muscovite (1%) and accessories (0.5%). Apart from the greater abundance of K-feldspar the texture and grain size of the other components strongly resembles that of the CBH 1 granodiorite On weathered surfaces the megacrystic K-feldspar have a chalky-white colour and gives the rock a much coarser appearance and also a slightly lighter grey colour than the CBH 1 granodiorite.

The CBH 2 monzogranite does not show any major variation in appearance apart from the ABC 13 area where it and CBH 3 monzogranite have a pronounced red discolouring, show a decrease in biotite content and are often intensely weathered. This N-S trending zone of discolouration appears to have formed by late stage brittle faulting and the possible influx of fluids.

CBH 3

This monzogranite is the dominant type within the Crobane Hill mapping area, occupying most of the north-eastern part of the map (A-D, 1-19). It contains as inclusions almost all of the previously mentioned granitoids, although in respect to the CBH 2 this relationship is less obvious as the observed enclosure of CBH 2 in CBH 3 may be a topographic effect. In hand specimen the CBH 3 monzogranite has a whitish-yellow colour and contains considerably less biotite than the other granitoids in this area. This monzogranite is aphyric with the grain size ranging between 3-5 mm. The mineral phases within CBH 3 are plagioclase (35.5%), K-feldspar (28%), quartz (31%), olive-green biotite (3.8%) muscovite (1.5%) and accessories (0.2%). CBH 3 has a low colour index giving it a light-grey colour on weathered surfaces and the absence of large K-feldspar megacrysts allows it to be distinguished from the CBH 2 porphyritic monzogranite. Throughout this mapping area CBH 3 shows no obvious grain size variation, but does have fluctuating amounts of biotite, especially in the zone of reddening mentioned in the previous section.

CBH 4

This biotite-rich microtonalite is the finest grained granitoid on Crobane Hill and commonly occurs as inclusions within the other granitoids. Despite occurring as xenoliths they preserve an overall linear trend oriented approximately parallel to the foliation within the pluton (see the central part of Map A). This feature may indicate, that originally, this microtonalite formed a once continuous dyke (see later) which has been subsequently deformed and dismembered, (analogous to the syn-plutonic dykes of Pitcher & Read (1960) and Pitcher & Berger (1972) (see later). In the squares A17-20 similar xenoliths are seen but do not display such a strong alignment. In hand specimen this tonalite is equigranular with an average grain size of less than 2.5 mm. It is composed chiefly of plagioclase (51%), quartz (31%), biotite (15%) with muscovite (1%), epidote (0.9%) and other accessories (0.2%). K-feldspar (< 1%) is rare to absent in CBH 4. The high biotite content, dark-grey colour and fine-grain size gives this granitoid a distinctive appearance on weathered surfaces, making it easily distinguishable from the more felsic CBH 2 and CBH 3 monzogranites.

Within the mapping area the CBH 4 tonalite shows very little variation in grain size or mineralogy. In some parts of the area though there are banded granitoids around some of these fine-grained masses (e.g. C7) implying the younger monzogranites (in this case CBH 3) have been deformed around older, solid masses of tonalite. This observation therefore suggests there maybe two compositionally similar tonalites which are of different ages but identical in appearance i.e. an earlier and later suite of microtonalites (see section 4:2:2). The older type of this tonalite, based on field relationships will be called CBH 4*.

CBH 5

This monzogranitic phase in the Crobane Hill mapping area is the volumetrically least important occurring as a large dyke, plus a series of smaller offshoots, traceable from B7 to C3 in the north-eastern part of the map. This dyke is up to 10 metres wide and maintains its regular width for a considerable length outside the area of detailed mapping. This dyke, plus the other smaller ones show little sign of disruption and usually have sharp, planar external contacts although occasionally lobate implying that the host may not have been fully crystallised at the time of emplacement. CBH 5 granite has a very similar grain size to the CBH 4 tonalite (i.e. < 3 mm), but usually contains generally less biotite giving this monzogranite a lighter grey appearance on weathered surfaces. Furthermore the finer grain size and equigranular texture allows it to be easily distinguished from the generally coarser CBH 1, CBH 2 and CBH 3 granitoids. The chief components are: quartz (35%), plagioclase (intensely serictized) (31%), K-feldspar (21%) biotite (9%), muscovite (3%) and epidote + other accessories (1%). Fresh surfaces reveal a brownish grey coloured granite in which biotite has a strongly preferred orientation producing a prominent foliation.

The CBH 5 monzogranite appears to correlate with the late microgranite dykes of Pitcher & Berger (1972), i.e. they have not been significantly deformed by subsequent movements in their host.

4:2:2 Field Relationships between the granitoids on Crobane Hill

In outcrop the contacts between the various granites are relatively sharp, whilst in thin-section the same contacts are difficult to recognise due to the interlocking nature of quartz and microcline crystals along these contact zones. The next section will describe field observations seen on outcrop scale within the mapping area which illustrate some of the important contact relationships between these respective granitoids.

❶ Figure 4:5 (a) and (b) shows the age relationship between a composite Thorr Granodiorite, pegmatite, and a fine-grained tonalite, CBH 4* (this fine tonalite appears to be early in age compared to the more typical CBH 4) xenolith in a host of CBH 3. This exposure is located at the northern end of B17. The oldest material is the Thorr granodiorite followed next by the pegmatite. The truncation of the pegmatite by both the CBH 4* and CBH 3 granitoids confirms this, although whether or not this pegmatite is related to the Thorr Pluton or the Main Donegal Pluton is uncertain at this location. It is clear that the CBH 4* tonalite is older than the CBH 3 monzogranite though its age relationship to CBH 1 is uncertain. The xenolithic nature of CBH 1 implies that it is older than CBH 3. The nature of the contacts between the respective granitoids in figure 4:3 will now be discussed, with an attempt to establish the rheologies between the various phases:-

i) *Thorr Granodiorite-CBH 4**:- this is a sharp, planar contact, clearly distinguishable due to grain size and colour contrast. Although sharp, close examination reveals that quartz crystals have grown along the contact, though the exact relationship is obscured by a 0.8 cm thick feldspar vein.

ii) *Thorr Granodiorite-CBH 3*:- the colour contrast clearly distinguishes the two (see photograph), although again coarse crystals of quartz can be seen across the contact. This contact is strongly oblique to the mineral alignment within the pluton. The interlocking nature of the weakly aligned quartz crystals with the Thorr contact implies that some of the deformation was syn-magmatic.

Figure 4:5:- (a) Photograph of a composite xenolith of Thorr Granodiorite, pegmatite, CBH 4* within a CBH 3 monzogranite host.

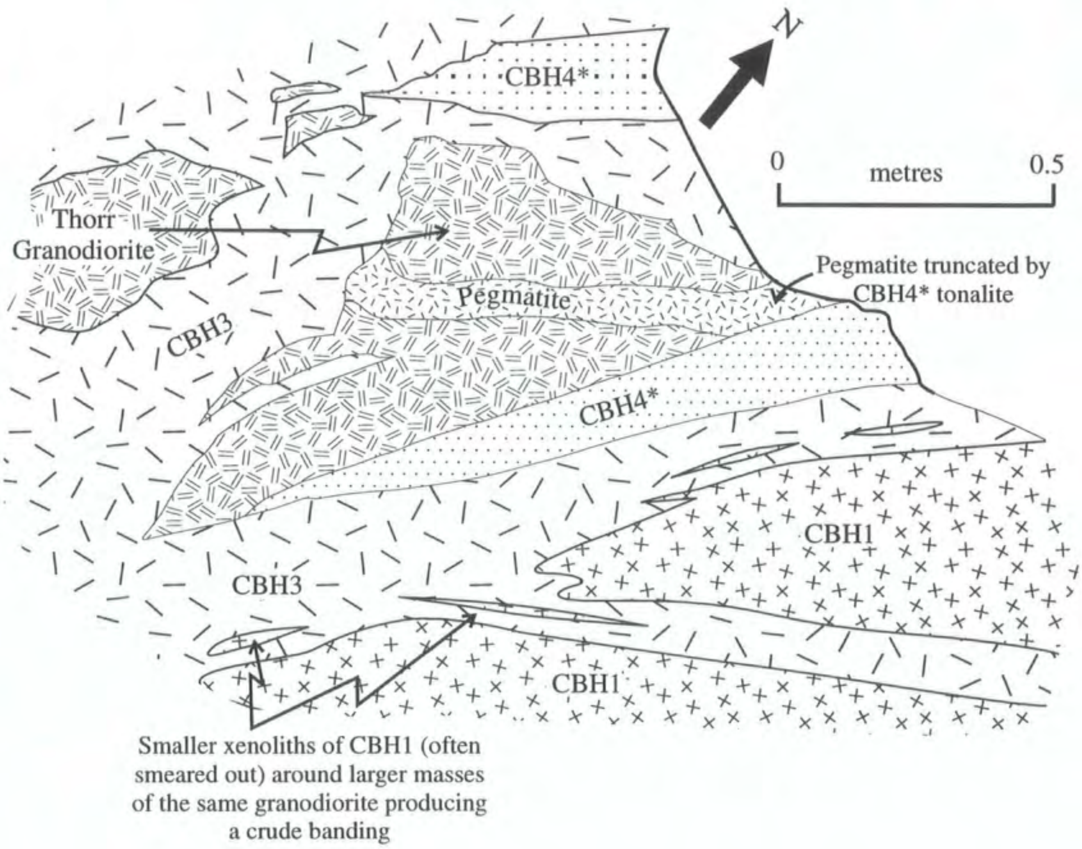


Figure 4:5:- (b) Sketch of (a) from a different angle illustrating cross-cutting relationships between various granitoid phases.

iii) *CBH 4*-CBH 3*:- The contact is highly planar and very sharp, although close examination reveals quartz crystals lying across the contact. The angular relationships between the CBH 3 monzogranite and the composite xenolith implies the latter was highly competent during the intrusion of the former.
iv) *CBH 1-CBH 3*:- Within the CBH 3 monzogranite there are faint wispy dark bands and biotitic schlieren, which may be have been produced by viscous flow of CBH 3 around CBH 1 granodioritic masses. This banding appears to have been produced by smaller fragments of CBH 1 spalling off the larger raft and subsequently thermally softened by the intruding CBH 3 monzogranite and smeared out during syn-plutonic deformation. Along the contacts between these two phases crystals of quartz and occasionally K-feldspar are clearly visible growing across it. This crude, irregular banding seen around CBH 1 masses occurs throughout the Crobane Hill mapping area and is strong evidence for this granite being relatively competent during the intrusion of the CBH 2 and CBH 3 monzogranites.

❷ As already mentioned there appears to be two generations of fine-grained tonalite. The most abundant tonalite form inclusions which preserve an overall linear trend (CBH 4) and are believed to belong to a dismembered dyke. The evidence for this is: (a) the presence of this tonalite in all the other granitoid types apart from CBH 5, and (b) the preservation of its linear trend. The other type of fine-grained tonalite appears to be earlier in age (CBH 4*), as indicated in the outcrop in ❶, predating CBH 3, although the exact relationship between it and CBH 1 and CBH 2 could not be ascertained from the mapping area. In figure 4:6 one can see a mass of CBH 4* surrounded by banded monzogranite which diminishes in intensity and eventually grades into unbanded CBH 3 monzogranite away from the mass. The banding is differentiated by the presence of darker, finer granite bands and coarser lighter bands containing large K-feldspar megacrysts (generally larger than that seen in the host CBH 3 monzogranite). This type of banding is almost identical to the regular banding seen at Doocharry (Berger 1971), further to the SW, although here the banding is developed on a much smaller scale. At Crobane Hill the exact mechanism of band formation is uncertain due to a lack of detailed petrographic and geochemical data, although the morphological similarities with the banding of Doocharry suggests a common cause. At Doocharry, Berger (1971) claimed the banding was produced by segregation of K-feldspar-rich fluids in to planar zones of movement during deformation of crystallising granitoid around solid masses of early formed tonalite. This suggests the width of banding around a raft is proportional to the size of the raft. In figure 4:6, the contact between CBH 4* "raft" and banded CBH 3 monzogranite is almost knife-sharp apart from the growth of quartz along the contact.

Figure 4:6
fine-grained
CBH 4* mass
with localised
banding
around it

Figure 4:7:-(a):- Localised banding along the contact between CBH 2 (left) and CBH 3 (right). These dark bands resemble CBH 1 in appearance.

Figure 4:7:-(b) Possible syn-magmatic shear separating CBH 2 (left) from CBH 3 (right).

Figure 4:7:- (c) Syn-magmatic shear from the Doocharry area where the host bands are more discordant to the foliation.

It therefore appears that within the Crobane Hill area there are two phases of tonalite. The CBH 4* definitely predates CBH 3 and possibly predates the CBH 1 and CBH 2 granitoids. The CBH 4* and CBH 4 tonalites are petrographically and geochemically similar and may form a suite of dykes which overlap with the intrusion of the major units, i.e. CBH 3.

❸ The age relationships between the CBH 2 and CBH 3 monzogranites is more uncertain due to the general absence of definite cross-cutting relationships between the two. As indicated in figure 4:7 (a) & (b) these two monzogranites are often separated by a form of granitic banding with the one grading progressively into the other. In figure 4:7 (a) (B17) the coarse-grained appearance of CBH 2 is very clear, whilst on the right is the pale-grey CBH 3 monzogranite. Separating the two is a 30 cm wide zone of banding of which the dark bands resemble the CBH 1 granodiorite. The presence of banding only in CBH 3 is rather equivocal evidence that CBH 2 might be older. Another alternative explanation is the contact has been a zone of movement within the cooling viscous CBH 2 and CBH 3 monzogranites with the CBH 1 type granodiorite having been smeared out along it. Figure 4:7 (b) (C17) provides stronger evidence for movement between CBH 2 and CBH 3 with this type of banding resembling that formed by syn-magmatic shearing, as seen elsewhere in the pluton. Figure 4:7 (c) shows a syn-magmatic shear from the Doocharry area for comparison ("cross band" of Berger 1971) where the deflection of older bands is more clear. The orientation of this banding at 060-240° on Crobane Hill is also consistent with other syn-magmatic shears seen elsewhere in the pluton. The upper age constraint on movement of this possible syn-magmatic shear is shown by the pegmatite vein which cross-cuts the syn-magmatic shear but shows no evidence of any disruption. Furthermore the absence of any intense foliation implies that movement had ceased before the formation of the mineral alignment.

Where there is no banding or syn-magmatic shears between the CBH 2 and CBH 3 monzogranites the contact is transitional over a distance of a few centimetres with the coarse K-feldspar megacrysts in CBH 2 gradually diminishing in size and frequency into the CBH 3 monzogranite. This implies that CBH 2 was not totally consolidated during the emplacement of the CBH 3. Therefore whilst the balance of evidence (including the fact that CBH 3 is more chemically evolved (see chapter 7)) favours the porphyritic CBH 2 being earlier than the aphyric CBH 3, there is no unequivocal relationship to establish this (see Sruhanavarnis section 4:3 for a similar scenario). Although equivocal the outcrop relationships in Map A imply the porphyritic CBH 2 monzogranite has been broken up by the aphyric CBH 3 monzogranite. Evidence for this conclusion is best seen in the south-western part of the mapping area (i.e. A-C, 16-19).

The CBH 1 granodiorite does not form any continuous exposure, instead it forms lensoidal masses with their long axes parallel to the foliation within the pluton. Along strike these masses eventually die out into schlieren-like bands (examples of this are best seen in the north-eastern part of the mapping area where exposure is almost 100%). As already mentioned there is usually a wispy banding developed around the CBH 1 masses, which often appears to be the result of smaller CBH 1 fragments entrained into the magma, becoming thermally softened by the intruding granitic material and then become flattened against the larger more competent CBH 1 mass. Figure 4:8 illustrates the lensoidal nature of the CBH 1 granodiorite with both ends tapering out into very thin bands. Angular contacts between CBH 1 and its host are occasionally seen at the termination of these raft-like masses, although usually the contacts are smooth and wavy.

Figure 4:8
Photograph illustrating the lensoidal nature of CBH 1 granodiorite rafts.

4:2:3 Deformation Structures within the Crobane Hill area

4:2:3:1 Outcrop

The strain within these granitoids is generally weak in this area, especially at outcrop scale. In outcrop there appeared to be no preferred alignment of plagioclase laths, although the presence of later microcline (deduced from thin-section textures) makes identification of early plagioclase fabrics more difficult in the field. On weathered surfaces there is usually a faint alignment of quartz, seen most easily in the equigranular granitoids, notably CBH 1. The sub-vertical foliation is consistently oriented in a NE-SW direction (see figure 4.9). No S-C' fabrics were encountered in this area.

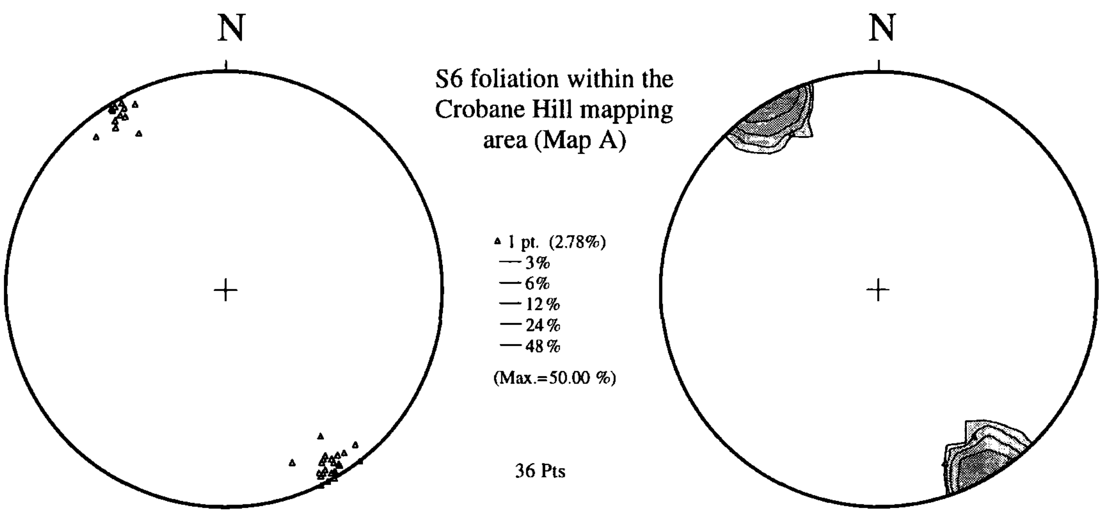


Figure 4:9:- equal area stereogram of foliation (S_6) in the granitoids on Crobane Hill.

At the NE end of the mapping the disrupted CBH 4 enclaves display an en echelon arrangement which may suggest sinistral shear (figure 4:10) A similar relationship is seen along much of the length of the CBH 4 "xenoliths" within the Crobane Hill mapping area.

4:2:3:2 Thin Section

Microscopic evidence of deformation is generally low within the majority of these granitoids with primary granitic textures preserved. There are no obvious PFC fabrics visible, apart from very faint alignments within the CBH 5 microgranite dyke. Within all of the granitoids there are moderate CPS fabrics. The behaviour of individual mineral phases within the granitoids of Crobane Hill will be described separately:-

1) *Plagioclase*:- tends to unaffected by the strain, with no evidence of any rounding. Albite twins are occasionally bent with some plagioclases showing "sweeping"

extinction. Passchier & Trouw (1996) stated this phenomena is commonly caused by sub-microscopic, brittle microfractures (visible only in TEM) rather than ductile deformation.

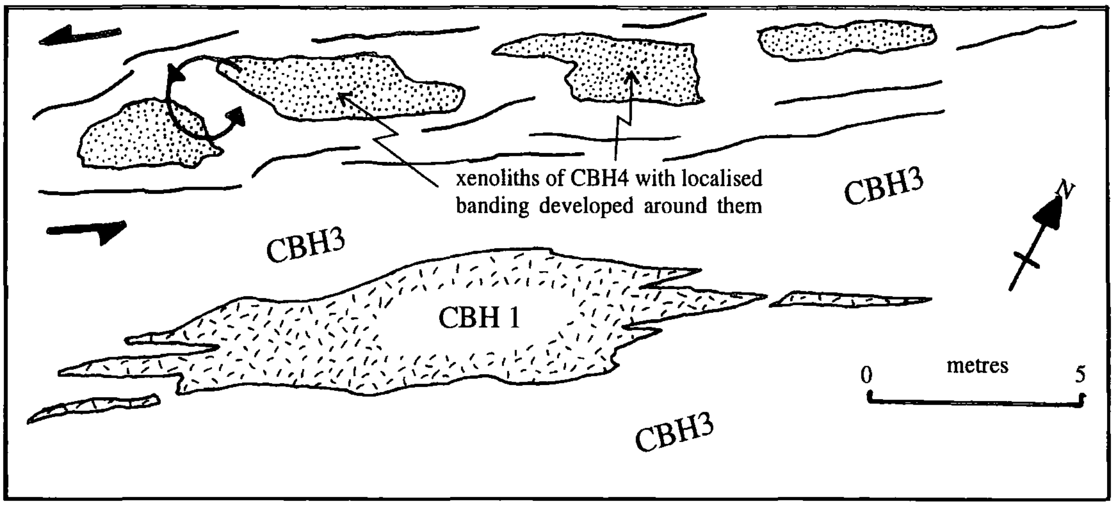


Figure 4:10:- field sketch showing the en echelon nature of the CBH 4 dismembered microtonalite dyke (100 metres NE of summit of Crobane Hill).

2) *K-feldspar*:- like plagioclase, this shows no evidence of having undergone any significant deformation. The majority of K-feldspar is in the form of microcline, although there is orthoclase which shows well developed "flame perthite". 3) *Quartz*:- this mineral displays the most evidence of deformation due to its higher susceptibility to strain. Most grains show well developed undulose extinction with localised development of subgrains. This suggests there is some degree of recovery by dislocation climb. Evidence for recovery by dynamic recrystallisation is minimal. The recovery by "climb" processes within quartz has allowed grains to become flattened without fracture (i.e. recovery is a working against strain hardening) producing a weak foliation within these granitoids. 4) *Biotite*:- strongly defines the foliation within these granitoids. There are occasional kinked biotites although overall they are essentially undeformed. A common occurrence is the presence of aligned biotite aggregates (internally undeformed) which suggests the new growth of biotite possibly by mimetic growth on an earlier aligned phase of biotite (Passchier & Trouw 1996). These earlier biotites tend to be chloritised and have irregular boundaries suggesting dissolution. The absence of internal deformation and the strong alignment within biotite may have been the product of syn-magmatic deformation, i.e. biotite PFC fabrics.

4:2:4 Summary of the chronology on Crobane Hill

The oldest granitic phase in this area is the Thorr Granodiorite which is presumed to have occupied this area prior to the emplacement of the Main Donegal pluton. The common occurrence of composite Thorr and Dalradian country-rocks xenoliths i.e. pelites, quartzites and metadolerites in the granitoids around Crobane Hill implies there was an extensive "ghost stratigraphy" within this earlier pluton. Whilst the full relationship of CBH 4* is not fully known it could be the earliest unit here, although the CBH 1 granodiorite is the demonstrably earlier phase in relation to the other major units (i.e. CBH 2 & CBH 3) . The original architecture of the CBH 1 prior to the emplacement of the later monzogranites is uncertain, although it appears that it was originally all one mass containing inclusions of Thorr pluton and metasediments. This granodiorite is now preserved only as rafts within the younger monzogranites with localised banding developed around these rafts. The alignment of these CBH 1 rafts may have been controlled by the emplacement of the later granitoids (i.e. they were fractured in this orientation by the intruding monzogranites) or they were fractured and then rotated into parallelism with the foliation (X-Y plane of the strain ellipsoid).

The CBH 2 monzogranite appears to be the next phase of intrusion, with outcrop relationships (in Map A) suggesting a sheet-like form now sub-parallel to the foliation within the pluton. The presence of some angular contacts between CBH 1 and CBH 2 implies that the former may have been relatively competent during the emplacement of the latter (i.e. displaying Bingham or visco-elastic properties). The CBH 3 aphyric monzogranite clearly has a sheet-like relationship and this disrupts all earlier granitoid sheets. The orientation of these sheets is again approximately NE-SW. The lack of angular contacts between CBH 2 and CBH 3 plus limited mixing and homogenisation along contacts implies both were viscous and capable of flow during intrusion.

The later microgranitoid phases only form minor intrusives in this area. The CBH 4 tonalite appears to be a dyke emplaced dominantly into CBH 3 (although it is in contact with CBH 1 as well as CBH 2). It is considered that these enclaves were originally a dyke which has been dismembered by subsequent deformation within the host, i.e. it has been boudinaged. This implies the CBH 3 monzogranite was less competent than the CBH 4 granite. The absence of any shear bands or fractures between the boudins implies the dyke was disrupted soon after intrusion and implies that CBH 3 was possibly capable of fracture at presumably high strain rates to allow the dyke to intrude but then flowed at lower strain rates allowing separation to occur, i.e. on average it displays visco-plastic or visco-elastic behaviour.

The later phase of minor intrusion within the mapping area is a series of microgranite dykes, which trend in a broadly E-W direction, of which the CBH 5 dyke is the most typical. This intrusion is very regular in width and the absence of deflection of markers (i.e. "drag" structures) within the host granitoids implies that the host was highly competent. The foliation within the CBH 5 body is generally much stronger than that seen in the host. The foliation is typically oriented between 065-070°, a few degrees clockwise of the foliation in the host. This obliquity of the foliation may imply that there has been dextral shear along the dyke. The orientation is compatible with that of anti-reidel, (R2) shearing within the pluton. Apart from the pegmatite-aplite vein there appears to be no other matchable markers across the dyke. This may suggest a considerable displacement across the 10m wide dyke and shear zone.

The final phase of intrusion on Crobane Hill was that of small regular pegmatite (sometimes composite aplite-pegmatite) veins. The lack of offset of such veins across the CBH 5 dyke implies the pegmatites are younger, although the lack of exposure in this critical area (i.e. C4) prevents this from being unequivocally proven. There appears to be two dominant directions; the more dominant set trending approximately N-S, whilst the other less common set is oriented E-W. The absence of displacements (revealed by markers) on either side of the pegmatites implies there was no lateral movement along these dykes and veins; instead they appear to be purely dilational in origin.

4:3 The SruhanavarnisValley

Introduction

Figure 4:11:- The Sruhanavarnis Valley (looking in a northwards direction towards Errigal).

The Sruhanavarnis valley lies in the heart of the Derryveagh Mountains, approximately 2.5 km ENE of Slieve Snaght. The valley is easily accessed from the R254 road (at GR B 957158) which passes along the Owenbarra valley and into which the Sruhanavarnis flows. Alternative access is also possible from the NW through the Poisoned Glen although this route is somewhat arduous.

A transect along the valley encounters the Glenveagh 3A raft-zone in the lower half (SE), whilst at the far (NW) end of the valley, at the "lip" of the Poisoned Glen the Derryveagh 2 raft-zone is seen. The granitoids between these two raft-zones are heterogeneous and beg similarities to those seen at Crobane Hill. Within the Sruhanavarnis detailed mapping (1:250) was performed on the granitoids that lie **within** the Glenveagh 3A raft-zone, whilst the north-western portion of the valley was only studied in a reconnaissance fashion. In this area the Glenveagh 3A raft-zone is up to 600-750 metres wide and is composed of rafts of varying composition, although the most typical components are pelite, calc-pelite, quartzite and meta-dolerite. In comparison to the Derryveagh 2 raft-zone the rafts are more discontinuous and tend to form "swarms" which preserve an overall linear trend parallel to the long axis of the pluton.

The area of detailed mapping was undertaken on the well exposed glacial pavement situated to the SW side of the Sruhanavarnis River forming a section which is 510 metres long and 90 metres wide and is almost perpendicular to the strike of the S_6 foliation, banding and the preferred orientation of the raft-trains (see figure 4:12 for general location). Three other isolated maps (60 $\times$ 90 metres) were compiled (a) in the stream bed of the Sruhanavarnis (b) 275 metres upstream of Lough Avarnis, and (c) 100 metres to the SW of Lough Attirive (big). (see figure 4:12).

The granitoids within the raft-trains are extremely heterogeneous, generally showing a higher degree of complexity, in regards to field relationships than those observed on Crobane Hill. In summary there are three main granitoid types encountered within Map B of this thesis:-

1) Fine-grained aphyric granodiorites.
2) Medium-grained aphyric tonalites.
3) Porphyritic monzogranites.

Within these basic divisions there are variations. Granitoids 1 and 2 have variable grain sizes and can be difficult to distinguish based on purely field criteria. Furthermore across strike in Map B the porphyritic monzogranites vary in appearance in regards to size of the groundmass, and to the size and colour of the megacrysts. Intruded into these granitoids are a suite of pegmatites-aplites and microgranite dykes. Within this area the granitoids have been sampled extensively for geochemical analysis in an attempt to ascertain whether or not any cryptic

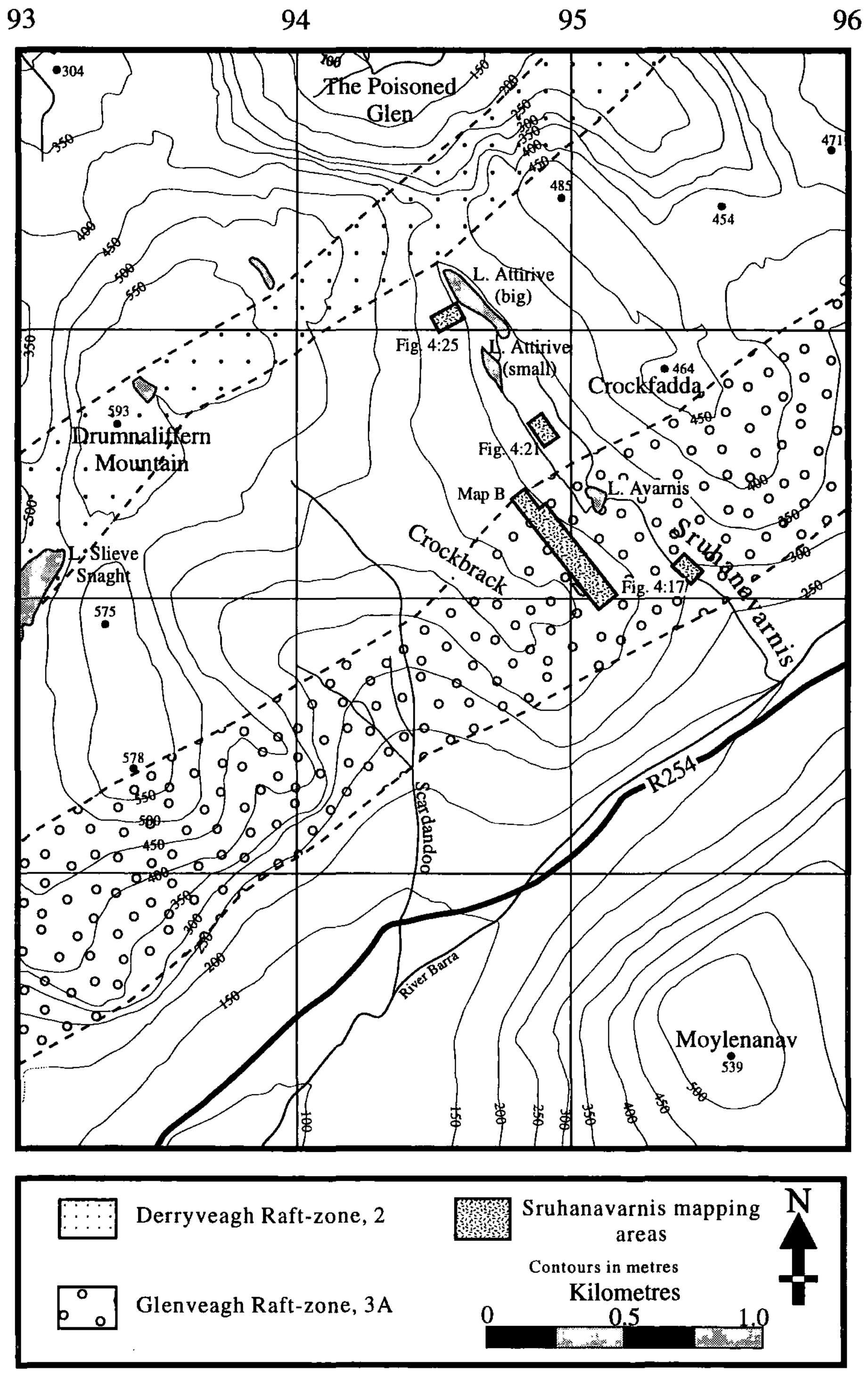


Figure 4:12:- map showing the location of the Sruhanavarnis mapping area and its relationship to major raft-zones.

variation exists. The results and interpretation of this will be addressed in chapter 7. Before going on to describe the granitoids within this mapping area a brief description of the meta-sedimentary raft-trains will be given.

4:3:1 Metasedimentary rafts within the Glenveagh 3A raft-zone

In general the rafts are aligned sub-parallel to the long axis of the pluton i.e. ~060°. Individual raft-trains have an en echelon arrangement (almost always right stepping), where when one train stops it can be seen to resume a few tens of metres across strike. The appearance of the raft-trains is as following:-

1) Pelite:- in this area the pelite rafts have been heavily permeated as shown in figure 4:13. Within the pelite rafts there are often veins of pegmatite and feldspar which appears to have been derived from assimilation of some of the pelite although one must emphasise that this only occurred on a very localised scale (0.5-1.0 metre width). Evidence of bedding is difficult to see due to strong transposition and pegmatisation. Pegmatite sheets within these rafts are intensely folded or boudinaged depending on their orientation with respects to the finite strain ellipsoid. In parts of the mapping area some of the pelite rafts, notably the smaller ones, have been almost totally assimilated, with their former existence marked by diffuse patches of pegmatite and biotitic schlieren.

In thin-section the pelite is composed of mainly biotite, microcline, quartz, muscovite and sodic plagioclase. The biotite forms the main foliation within the pelite and contains abundant equant zircons which have distinct pleochroic haloes. Microcline has been heavily impregnated by myrmekitic growths. Muscovite, which appears to have grown later than some of the biotite, is often kinked and folded and commonly contains opaque inclusions. Quartz displays varying degrees of undulose extinction, with there being evidence of dynamic recrystallisation in the weakly deformed crystals of quartz. Feldspars show evidence of later fracturing with the cracks infilled with iron oxides (see photomicrograph in figure 4:14) Within the pelite samples collected no sillimanite or fibrolite was encountered although previous workers have documented its presence (e.g. Pitcher & Read 1959).

2) Meta-dolerite: this tends to form massive blocks which show very little interaction with the host granitoids. In thin-section the meta-dolerite is composed of green pleochroic hornblende which shows a moderate to strong alignment; quartz, plagioclase (oligoclase-andesine), biotite, relict opaque iron oxides and in some samples garnet. The biotite, although not abundant, appears to have grown by replacement of some of the hornblende. Opaque iron oxides often possess relict inclusions of quartz and hornblende which preserve an internal fabric which is oblique to the weak external fabric. Plagioclase shows varying degrees of undulose extinction implying dislocation glide may be occurring within this mineral phase. Quartz is relatively undeformed but shows evidence of extensive strain healing with recrystallised grains having 120° triple junctions (see photomicrograph: figure 4:15).

Figure 4:13:- Heavily permeated nature to the pelitic rafts within the Sruhanavarnis Valley.

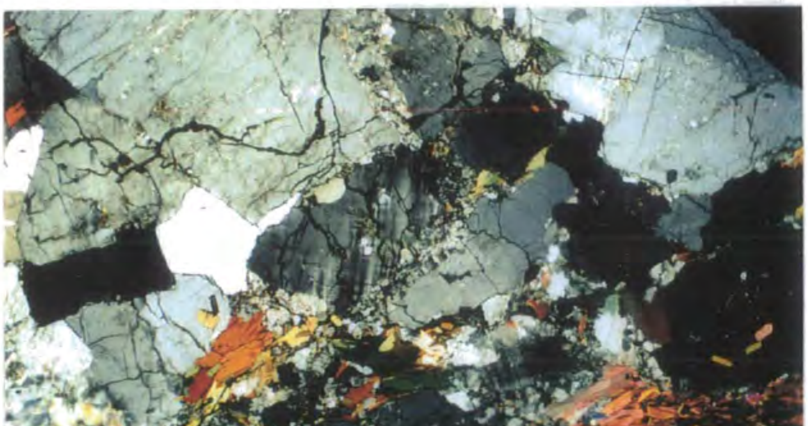


Figure 4:14:- Photomicrograph of pelite. (scale bar = 1mm). High temperature deformation overprinted by low-temperature deformation (note fracturing of feldspars).

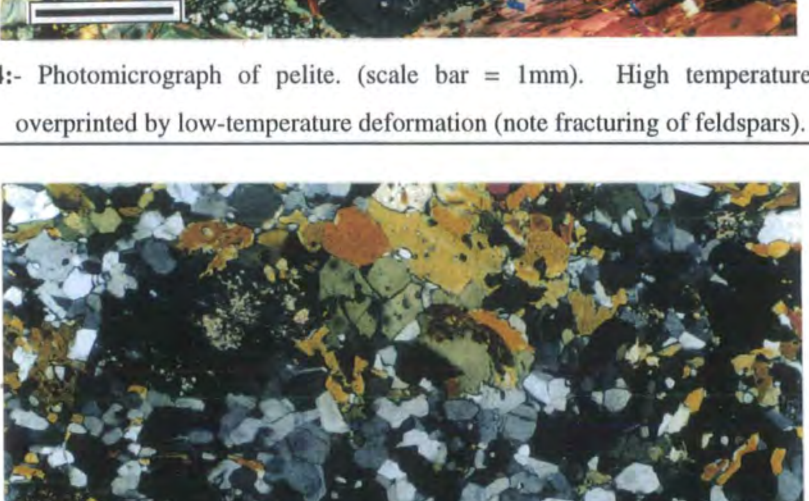


Figure 4:15:- Photomicrograph of metadolerite rafts which show high-temperature deformation (quartz is extensively recrystallised) (scale bar = 0.5mm).

Apart from the presence of garnet the meta-dolerite enclaves within these central regions of the pluton have predominantly nematoblastic textures. The behaviour of quartz and feldspar implies these metadolerites have been subjected to high temperature deformation, possibly prior, or during the emplacement of the Main Donegal Granite.

3) Quartzites:- these enclaves show the least sign of alteration with original bedding often preserved as dark mineral bands. Within these rafts the granotoids have often exploited the flagginess resulting in the development of small rafts aligned parallel to the foliation. In thin-section the texture is dominantly granoblastic with quartz grains up to 2-3 mm in size. Other minerals which only comprise a very small percentage of the rock, are microcline, plagioclase (heavily altered) and muscovite, the latter is aligned to produce a low degree of flagginess within the quartzite. The quartz shows only weak undulose extinction forming slightly elongate grains (parallel to the aligned muscovite). Dynamic recrystallisation has occurred to a limited degree by SR and GBM processes. The presence of straight crystal contacts and 120° triple junctions implies there has been limited secondary recrystallisation, probably after moderately high plastic strains had ceased.

4:3:2 Granitoid facies within the Sruhanavarnis Valley

The following section will address the dominant granitoid phases which are present in the Sruhanavarnis valley, especially those within Map B. The granitoids of this area will be prefixed "SRU" to avoid confusion with the granitoids described at Crobane Hill (figure 4:16 shows the modal classification of these granitoids). Photographs of the polished hand specimens are shown in Appendix A.

SRU 1

Of the main phases the SRU 1 granodiorite is volumetrically the least abundant tending to occur as rafts within all of the other granitoid types. This granodiorite also tends to be distributed in close vicinity or in contact with meta-sedimentary rafts which implies it was the earliest granitoid to intrude this area (and possibly of the entire Main Donegal pluton) (see section 4:3:6). In hand specimen this granodiorite is a fine to medium-grained (grain size typically 1-2 mm), composed of white plagioclase (42%), grey quartz (21%), white K-feldspar (20.5%), biotite (14%), muscovite (1.5%) with epidote (0.5%) forming the main accessory. The high mafic component (average 16.5%) gives this rock a high colour index. On weathered surfaces the fine-grained nature and dark-grey appearance allows it, in most cases, to be distinguished from the other granitoids. Another factor which allows it to identified on weathered surfaces is the growth of a green algae. It is apparent that this flora has strong affinities for elements in the mafic minerals (i.e. Mg^{2+} and $Fe^{2+/3+}$) and does not grow on the later, less mafic granitoid facies.

This granodiorite does show some variation, mainly in grain size which occasionally makes its appearance similar to that of the SRU 2 tonalite. The presence or absence of K-feldspar is another factor, but a variation only clearly identifiable in thin-section.

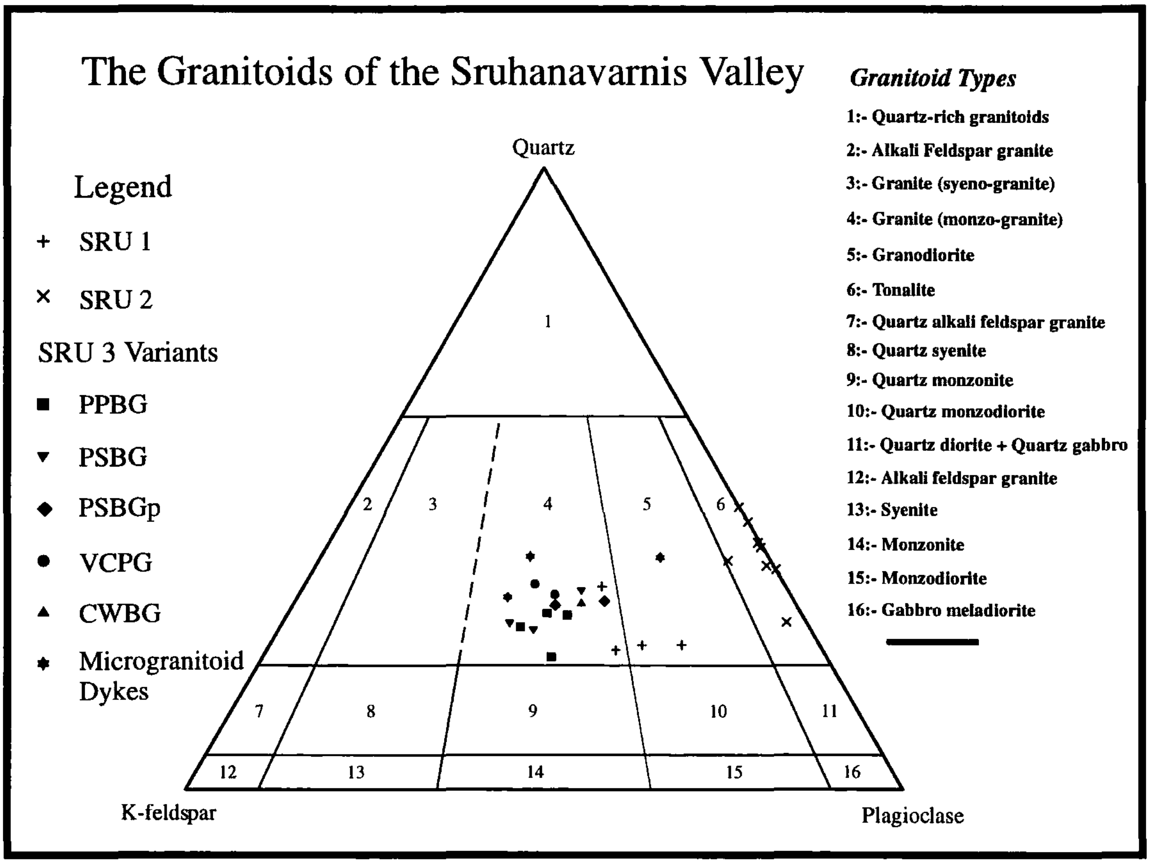


Figure 4:16:- QAP plot for the granitoids in the Sruhanavarnis valley.

SRU 2

This tonalitic phase has a more widespread distribution than the SRU 1 granodiorite and again appears to behave as rafts within the younger porphyritic monzogranites. Generally this rock is a medium-grained, equigranular tonalite-trondhjemite (silica-rich tonalite) composed of white plagioclase (51%), grey quartz (32%), biotite (14%) and accessory muscovite 1.5%) with trace amounts of epidote (0.25%) and apatite (0.25%). K-feldspar is rare to absent (~1%). The grain size of this tonalite is typically 2-4 mm and hence gives this rock a slightly lighter grey colour in comparison to SRU 1 due to the larger size of the plagioclase. In some areas of Map B this tonalite forms quite large exposures whilst in others it has been disrupted by subsequent monzogranites to produce a crude form of banding which is broadly parallel to the foliation (see sub-map 5 of Map B).

SRU 3

Apart from the pegmatites and the late microgranite dykes these porphyritic monzogranites form the youngest and most abundant phase tending to have large inclusions of the two earlier granitoid phases, already described, within it. Also it is the most common type in the Sruhanavarnis mapping area, although its appearance does alter along the length of the valley. The variation is most obvious in hand specimen, notably fresh surfaces (see appendix A) with the mineralogical differences only very minor. The average composition is:- plagioclase (34%), K-feldspar (31.5%), quartz (26%), biotite (7%), muscovite (1.25%) and epidote, apatite and opaque oxides forming accessory (0.25%) All of these SRU 3 variants plot within the monzogranite domain of the QAP plot where the proportion of plagioclase and K-feldspar is approximately equal (see figure 4:16) with the mafic content ranging from 5.3 - 12.6%. The main features of these variants in hand specimen will be addressed and accompanied with their overall distribution in the Sruhanavarnis valley:-

i) **PSBG**:- the finest grained porphyritic monzogranite which has a groundmass of 1-2 mm with megacrysts of white K-feldspar measuring up to 7 mm in diameter. This monzogranite forms the majority of the host in sub-maps 1, 2, 3, 4 and most of 5, although it contains a belt of a coarser variety (VCPG). Overall the texture and composition of the groundmass is similar to that of the SRU 2 tonalite.

ii) **VCPG**:- the coarsest of the SRU 3 monzogranites, forming a belt up to 40-90 metres wide, creates the highest topography in the Map B area. The grain size of the groundmass is 3-4 mm and the rock contains white K-feldspar megacrysts up to 13 mm in diameter. The grain size of the latter gives SRU 3: VCPG an obvious appearance on weathered surfaces.

iii) **PSBGp**:- essentially the same as i) but contains pinkish megacrysts of K-feldspar, although the groundmass is slightly finer grained. The absence of pinkness in other mineral phases such as plagioclase implies the pink colour is a primary feature and not the result of deformation assisted fluid flow or the close proximity of Tertiary basalt dykes. This monzogranite forms narrow belts within sub-maps 6-8 and lies overall within the PPBG granite.

iv) **PPBG**:- the groundmass of this variety is generally coarser (2-3 mm) than that of PSBG and PSBGp and contains pink megacrysts of K-feldspar (8-10 mm in diameter) which are generally in higher abundance compared to the PSBGp monzogranite. In comparison with PSBG it generally contains less biotite. This monzogranite is the main porphyritic phase in sub-maps 5 & 8 where rafts of quartzite, meta-dolerite, and to a lesser extent, pelite are abundant.

v) CWBG:- one of the coarsest of the SRU 3 variants seen at the "head" of the valley (to the NW of Lough Attirive (little). The groundmass ranges from 3-6 mm, whilst K-feldspar ranges from 6-12 mm with this latter mineral having a pinkish tinge. The coarser appearance of CWBG in hand specimen makes it the most distinct member of the SRU 3 porphyritic monzogranites.

Microgranitoid dykes and minor intrusions

Along the Sruhanavarnis section there are a series of microgranitoid dykes often traceable for considerable distances along strike and tending to maintain consistent widths. This is in contrast to the pegmatites which are generally smaller, more abundant and less continuous along strike. Sub-map 9 (figure 4:17) shows an early microtonalite dyke which appears syn-plutonic (figure 4:19). This dyke was described in ample detail by Pitcher & Read (1960). More common are the large microgranite-microgranodioritic dykes which strongly resemble those seen in the Crobane Hill section (i.e. CBH 5). These generally strike sub-parallel to the foliation and are broadly concordant with the strike of the rafts. The two large dykes in the section (one in sub-maps 3 & 4 and the other in map 8) both have consistent thicknesses of 10-12 metres and can be traced for up to 400-600 metres along strike outside the area of the map. In hand specimen these dykes are fine-grained, (<2 mm grain size) aphyric and tend to be strongly foliated. In thin-section these granites contain zoned plagioclase, (intensely serticised) (31.5%), quartz (31.5%), microcline (26.5%), olive green biotite (9%), muscovite (1.25%) and accessory epidote (0.25%).

Pegmatites:- present as two main types. The first type form small diffuse pods which are generally seen in close vicinity to pelitic xenoliths, for example in squares J3 (sub-map 4) O1 (sub-map 2). Often these pegmatite pods contain biotite schlieren (oriented parallel to the foliation, i.e. 058°) and are probably produced by localised assimilation of pelitic material. The more common pegmatites tend to form small regular veins ranging in size from 7-75 cm. The orientation of the pegmatites within Map B were plotted on a rose plot to see if there is any structural control on their distribution, i.e. joint planes etc. Figure 4:18 shows that the majority of these dykes are oriented NW-SE (~148°) although there is a lesser developed set at almost right anglesto this set. The strong alignment implies there may have been an early joint system within the cooling pluton which behaved as planar zones of weakness. The value of 148° is in general accordance with that of Pitcher & Read (1959) who stated throughout most of the pluton pegmatites and aplites were commonly oriented at

150°. Evidence from cross-cutting relationships shows the more regular pegmatites are younger and often truncate the more diffuse pegmatite pods, e.g. P1 of sub-map 1.

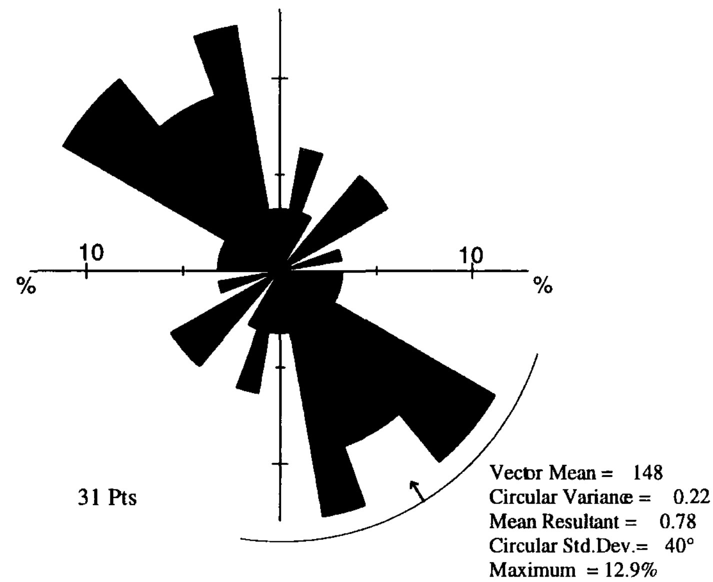


Figure 4:18
Frequency-azimuth rose plot of the trends of pegmatite and microgranite dykes, Sruhanavarnis valley.

4:3:3 Field Relationships between the granitoids of the Sruhanavarnis Valley

This section describes the contact and general field relationships between the respective granitoids. The contacts between these different phases were carefully examined to determine if they have any characteristic features, i.e.:- i) sharp ii) welded contacts. iii) contacts which have feldspar and quartz situated along them. iii) transitional contacts (over a distance of 0.1-1.0 metres). Although sound in theory, when applied to the field it was often more problematical due to the quality of exposure and preferential erosion along contacts preventing observation of the exact relationships from being seen. In the central area of Map B the precise nature of the contacts between the various granitoids were very subtle and almost impossible to see. The relationships become much clearer when the sun was shining on wet surfaces after a heavy rain shower. In Appendix B photocopies of the annotated field slips are included which describes these field relationships.

Within the mapping area contacts between SRU 1 and SRU 2 are relatively rare, with the age relationship only determinable at a few localities. In map B (O2) xenoliths of SRU 1 are found within SRU 2. The long axis of these xenoliths are aligned parallel to the foliation. Around larger inclusions of SRU 1 granodiorite there is a type of banding produced by smearing out of smaller fragments against larger ones. Contacts are generally sharp although often welded with quartz crystals across the contacts. In no areas were chilled margins observed in the SRU 2 tonalite where juxtaposed against SRU 1.

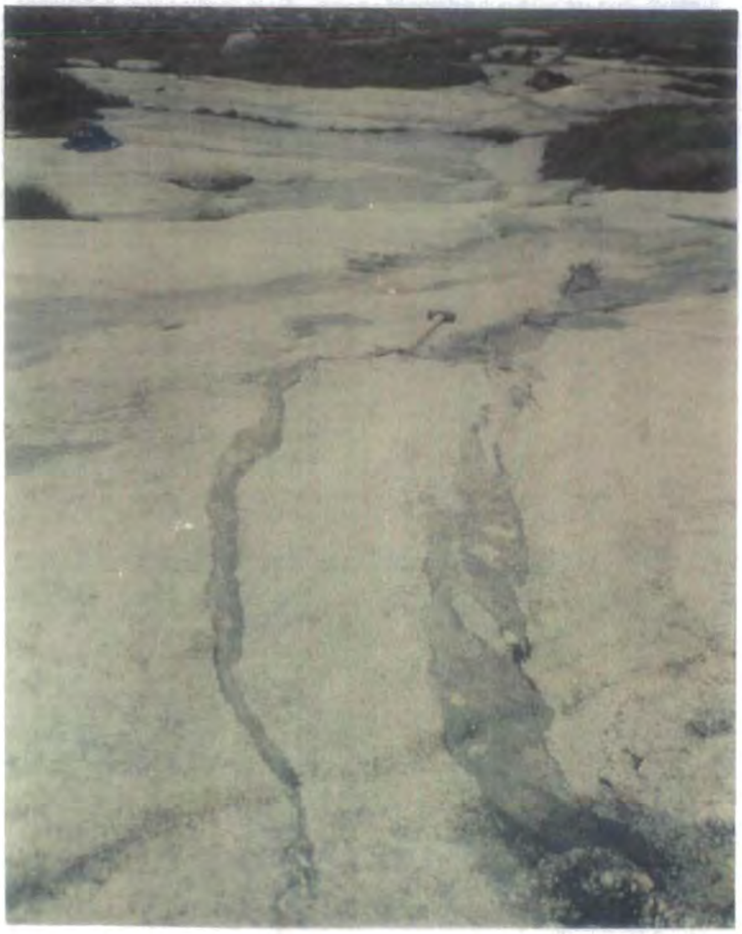


Figure 4:19
Syn-plutonic micro-tonalite within sub-map 9 (fig. 4:17) which shows signs of disruption by subsequent movements in the host. (The early transverse dyke of Pitcher & Read 1960a)

Figure 4:20:- The breaking-up of relatively competent SRU 1 granodiorite by the SRU 3 porphyritic monzogranites.

Contacts between SRU 1 and SRU 3 are sharp in outcrop scale although in thin section they are often interlocking. SRU 1 often displays angular contacts in the coarser porphyritic monzogranites (see figure 4:20) implying the former was relatively competent during the intrusion of the latter. In areas of poorer quality exposure the contacts are very faint to absent with the finer grained granodiorite demarcated by growths of green algae on their surfaces.

The lower contrast in grain size between SRU 2 and SRU 3 makes these contacts generally more subtle in appearance particularly where the quality of exposure is poor. Contacts observed commonly show both quartz and K-feldspar growing across the contacts.

As already described in section 4:3:2:2 there is variation within the SRU 3 monzogranite, which may indicate that there are separate pulses. Within the Sruhanavarnis mapping area there are no outcrop scale contacts visible between these monzogranites. The change in variety is based on careful comparison of fresh hand specimens and rock cores drilled across the strike within Map B. Contamination due to country rock assimilation can be ruled out as the porphyritic granites in the close vicinity of the rafts are identical to those collected further away from such rafts (see chapter 7, section 7:4:1:1). On the whole weathered surfaces are very similar. Key areas were studied where transitions between these monzogranites were seen but often it was found that the older SRU 1 and SRU 2 granitoids separated the SRU 3 variants.

4:3:3 The granitoids at the NW end of the Sruhanavarnis Valley

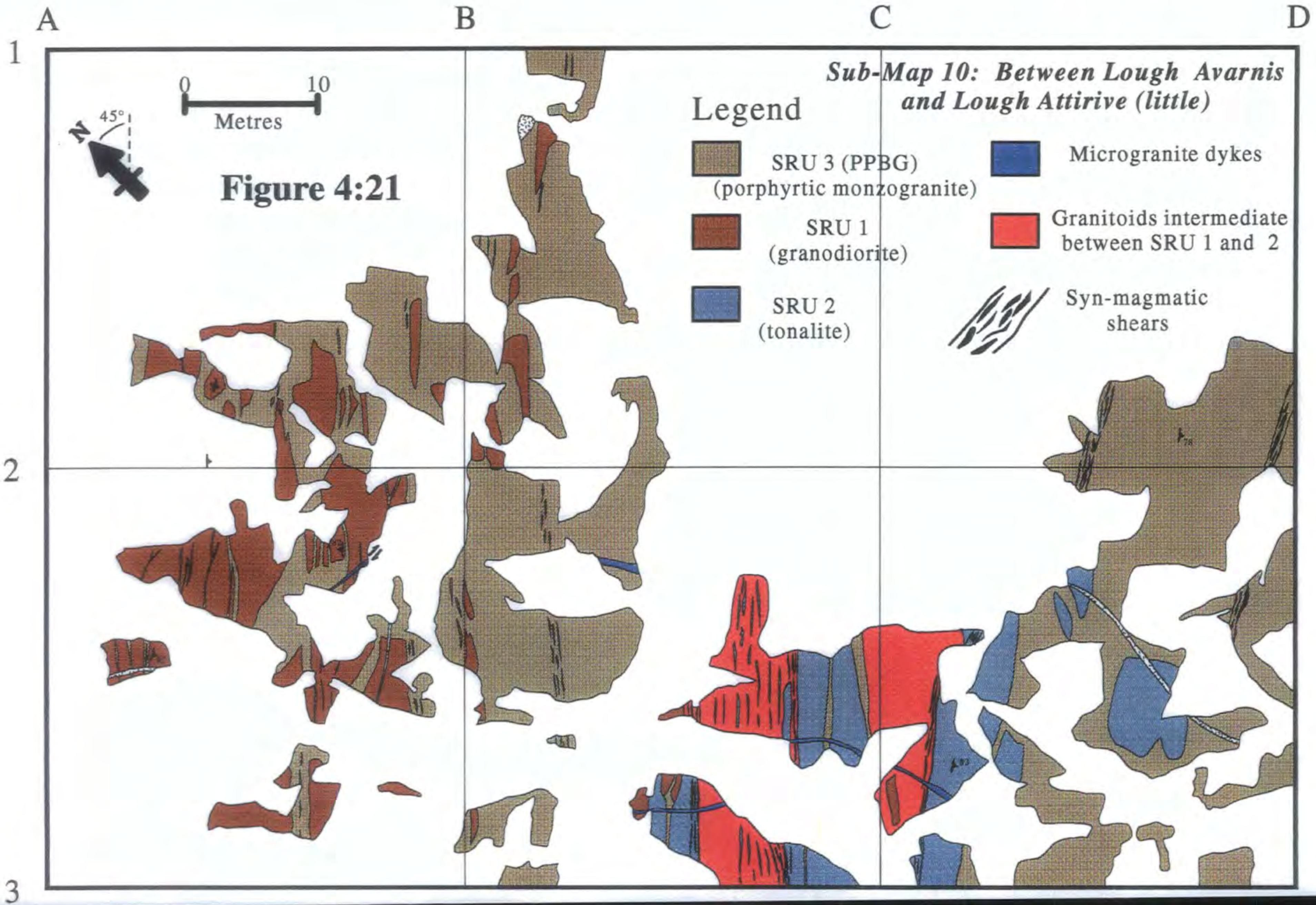
Despite the percentage of exposure being generally less than within the area covered by Map B the quality of exposure is higher due to a combination of peat bog "bleaching" and the presence of running water which keeps exposures looking clean.

Despite raft-trains being absent (apart from the large pelite raft (9 × 125 m) at GR 952159) a combination of fine-grained and medium-grained equigranular granitoids are present which resemble SRU 1 and SRU 2 respectively. These granitoids are in a host of variable SRU 3 porphyritic monzogranites. The overall distribution of the granitoids will be discussed traversing from the SE to the NW. Two detailed maps (90 × 60 metres) were made in this zone (see map 4:12 for location). Conclusions from Sub-map 10 (figure 4:21) are:-

a) SRU 1 type granodiorite is more abundant just outside the raft-zone than within it.

b) the host to SRU 1 is a pinkish porphyritic monzogranitic host which strongly resembles SRU 3: PPBG seen in Map B.

c) Small linear belts of banding probably represent syn-magmatic shears



A
B
C
D
1
2
3
Sub-Map 10: Between Lough Avarnis and Lough Attirive (little)
0
10
Metres
45°
N
Figure 4:21
Legend
SRU 3 (PPBG) (porphyrtic monzogranite)
SRU 1 (granodiorite)
SRU 2 (tonalite)
Microgranite dykes
Granitoids intermediate between SRU 1 and 2
Syn-magmatic shears

Figure 4:22:- (a) Sheets of SRU 3 porphyritic monzogranite intruding through the SRU 1 granodiorite (sub-map 10: figure 4:21).

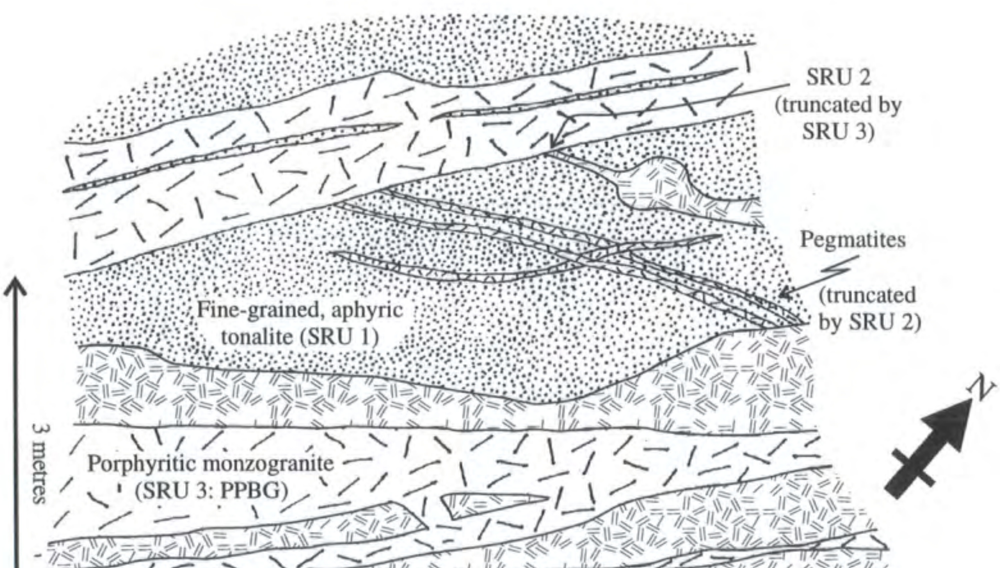


Figure 4:22:- (b) Field sketch showing cross-cutting relationships between SRU 1 granodiorite, SRU 2 tonalite and the SRU 3 (PPBG) monzogranite.

Figure 4:23:- (a) SRU 1 type granodiorite intruded by SRU 3 (PPBG) monzogranite. The irregular contact suggests the former may not have been fully competent (see text) during the intrusion of the latter.

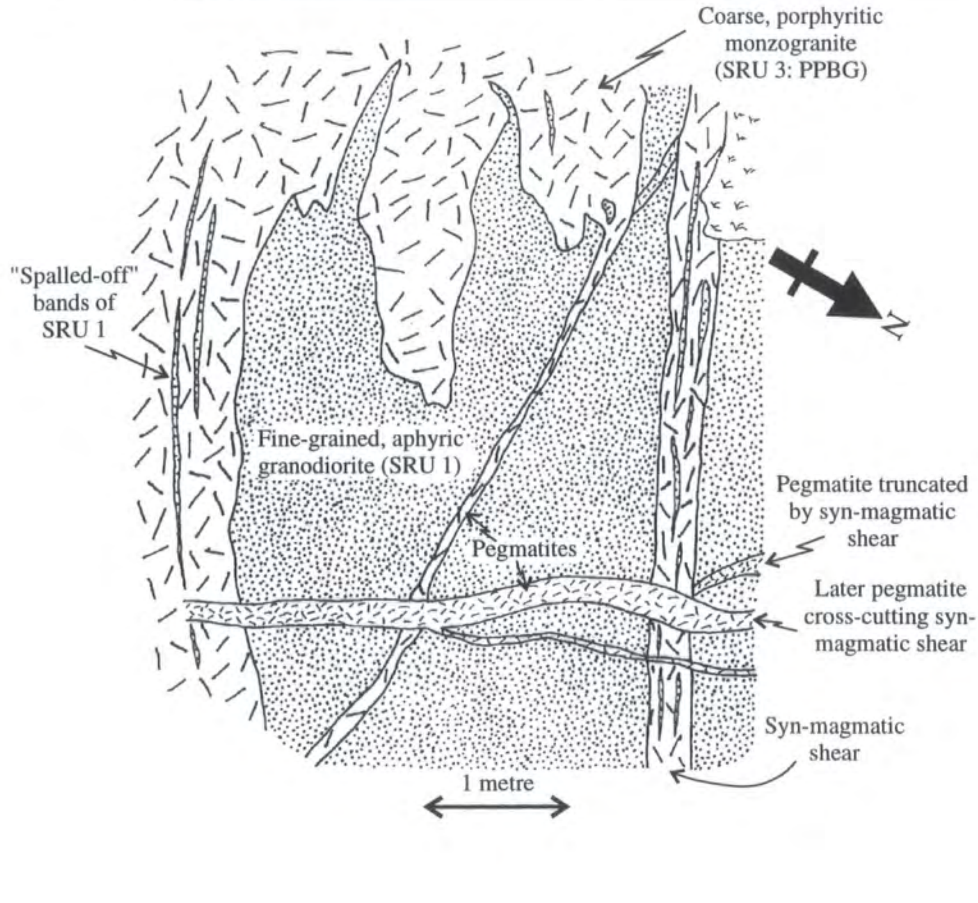


Figure 4:23:- (b) Field sketch of the photograph in (a)

This style of intrusion where the SRU 1 granodiorite has been broken up in to rafts by the younger porphyritic monzogranite continues for 275 metres to the NW of Map B (see figures 4:22 a & b and figures 4:23 a & b). In figure 4:22b the granodiorites and tonalites are highly competent with both these granitoids having sharp contacts with the SRU 3. The truncation of the pegmatite vein within SRU 1 by SRU 2 also implies the former was highly competent during the intrusion of the latter. In contrast, in figure 4:23 a & b, the more lobate contacts between the SRU 1 type granodiorite and SRU 3 monzogranite (PPBG) implies the viscosity contrast was much lower (i.e. SRU 1 was not fully crytsallised). This either suggest some of the earlier granodioritic material was still quite viscous during intrusion of SRU 3 *or* this granodiorite belongs to later phase of dyke intrusion which was subsequently disrupted after emplacement.

At GR B 946156 the SRU 2 tonalite becomes more abundant as the percentage of SRU 1 decreases, although the host SRU 3 monzogranite is still similar to the PPBG variant. Homogeneous masses of SRU 2 are commonly seen intruded by veins of porphyritic monzogranite between 0.5-0.75 metres wide. Banding between these two granitoids becomes quite common notably on the SW shores of Lough Attirive (little). A spectacular example is shown in figure 4:24 where parallel "strips" of SRU 2 (picked out by algal growths on surfaces) have been broken up by PPBG. This produces a crude type of banding parallel to the foliation.

Figure 4:24:- Granitic banding formed by strips of SRU 2 tonalite which have broken apart during the intrusion of the SRU 3: PPBG monzogranite (algal growths cover the tonalite bands).

At GR B 946159 the variety of SRU 3 changes with the PPBG host monzogranite disappearing and the somewhat coarser and more biotitic CWBG monzogranite now forming the host. The exact transition between these two variants is uncertain due to the relatively poor exposure but it is believed that this may be some form of major sheet contact between the SRU 3 variants. The emplacement style in this north-western part of the valley is typified by large homogeneous masses (rafts) of SRU 2 (and possibly SRU 1) tonalite within CWBG, the latter of which, is in places is very homogeneous (figure 4:25 typifies the style of emplacement at this north-eastern end of the valley). Generally, the granitoids at the far north-western part of the valley tend to be more homogeneous, with banding less common, than compared to the granitoids observed to the SE of Lough Attirive (little). A point of interest though are the increasingly common late intrusive dykes of fine-grained granodiorite which in hand specimen resemble the SRU 1 granodiorite (see figure 4:25). These dykes range from 0.5-10 metres in width and are traceable for considerable distances along strike. The dykes show varying degrees of disruption by subsequent movement in the host. Some dykes along parts of their lengths have been broken up into sub-angular inclusions which preserve the overall trend of the dyke. It appears that such dykes have been boudinaged with there being a very high ductility contrast with the dyke essentially deforming in a brittle fashion, at least at the outcrop scale.

Deformation within the granitoids of this area increases towards the NW end of the valley with the development of S-C' fabrics which show predominantly sinistral offsets, with the spacing of the "C" planes between 4-5 mm and may relate to high strains localising within the nearby Derryveagh 2 raft-zone. At the northern end of Lough Attirive (big), GR B 955162 (close to a small cave and the "wind gap" (see figure 4:12)) quartzite, plus occasional intercalations of pelite rafts are encountered. These rafts of quartzite form a belt 25-35 metres wide and have been intruded by a network of pegmatitic veins. To the NW of the quartzite there are xenoliths of Thorr Granodiorite which are often angular and form a zone approximately 175 metres wide. The rafts of quartzite and Thorr Granodiorite comprise the Derryveagh 2 raft-zone which on the whole is narrower than the Glenveagh 3A raft-zone but is laterally more continuous. The north-western boundary of this raft-zone is very sharp and abrupt and gives way to regularly banded monzogranite. At the immediate boundary the xenoliths of Thorr are intensely smeared-out parallel to the foliation within the granite implying that this raft-zone may have localised high strains. At Lough Maumbeg (GR B 938162) the regular banding is well developed forming a zone 120-150 metes wide, to the NW of which are very homogeneous pink porphyritic monzogranites (see chapter 5, section 5:3). The significance of this raft-zone will addressed in the following two chapters. This regular banding is not seen on the south-eastern side of this raft-zone.

4:3:4 Granitic Banding

Within the Sruhanavarnis valley there are many spectacular examples of banding which have been produced in a variety of ways:-

1) by the intrusion of fine-grained microgranitoid dykes which are parallel to the foliation.

2) by the breaking up of older granitoid into strips which are aligned parallel the foliation. The general lack of intense deformation within these older rafts implies they were not transposed during solid-state deformation but that they were possibly fractured in this orientation by the later SRU 3 type porphyritic monzogranites (figure 4:24). N.B. some rotation of these fractured rafts may have occurred when SRU 3 was still capable of viscous flow.

3) "Schlieren"-type banding: the presence of banding around rafts of countryrock or earlier formed granitic material. The bands are commonly composed of biotite rich granites, similar in composition to the masses which they are being deformed around, and show evidence of deformation, by possible viscous flow around a more competent mass.

4) Syn-magmatic shearing:- these appear as narrow zones (0.2-0.5 metres) of porphyritic monzogranite irregularly banded with finer grained more biotitic granite. These structures are the same as the "cross-bands" Berger (1971) observed in the Doocharry area. Furthermore the orientations of the cross-bands (syn-magmatic shears) are generally similar trending parallel or sub-parallel with the foliation in the pluton. In the Sruhanavarnis the recognition of this feature is not as obvious due to the lower abundance of markers discordant to the foliation. Figure 4:26 shows two examples of syn-magmatic shears within the Sruhanavarnis Valley, one of which (4:26a) displaces discordant banding in a sinistral fashion.

4:3:5 Deformation structures within the Sruhanavarnis Valley

4:3:5:1 Outcrop

Overall the strain within the granitoids in the Sruhanavarnis Valley is relatively low. Very weak plagioclase PFC fabrics have been observed in thin sections of the SRU 2 tonalite with the alignment broadly parallel to the S_6 foliation (see figure 4:27). The presence of numerous localised syn-magmatic shears attest to deformation occurring within the magmatic to sub-magmatic state.

Boudinaged granite veins within quartzite and pelite rafts:- within many of the meta-sedimentary rafts there are veins of pegmatitic granite which have been boudinaged

Figure 4:26:-(a) Discrete syn-magmatic shears displacing discordant tonalite bands in a sinistral manner.

Figure 4:26:- (b) Syn-magmatic shear in sub-map 10 where no markers are present to show any displacement.

e.g. sub-map 3 (L1), sub-map 8 (B2) of Map B and sub-map 9 (figure 4:17). This relationship implies the granitic veins were more competent than its host, implying that the pluton was being deformed quite late into its cooling history. Within the necks of the boudins there are shear bands, which although sometimes conjugate usually have the sinistral set better developed. This implies there were strong components of both flattening and sinistral shear acting upon the pluton (see figures 4:28 a,b & c). It is the authors belief that these raft-zones may have localised strain during the cooling of the entire pluton. Within the metadolerites there is evidence of high temperature deformation, whilst in the pelites (figure 4:14) there is evidence of down-temperature deformation with crystal plastic fabrics being overprinted by brittle microfracturing.

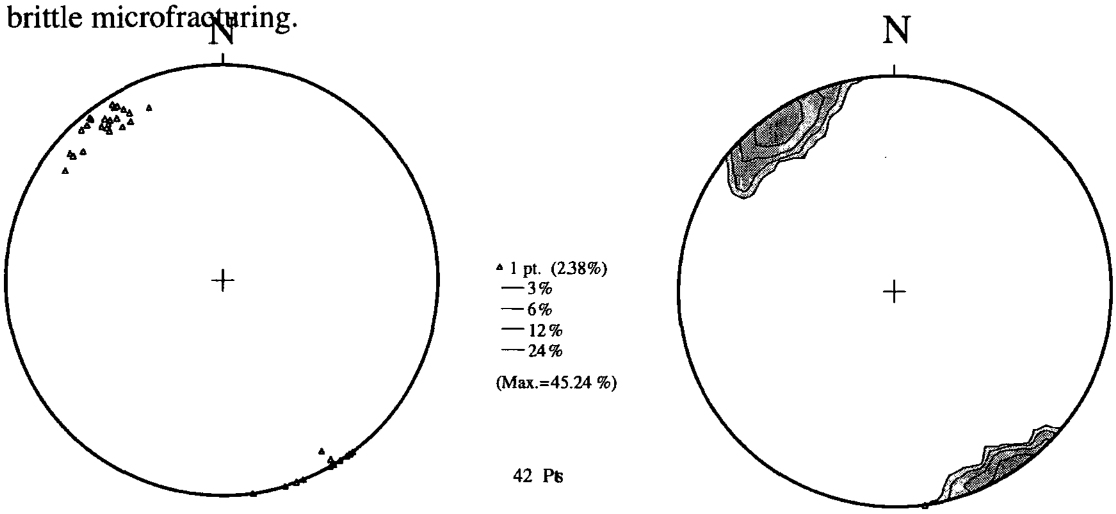


Figure 4:27:- Stereogram showing orientation of the S_6 foliation within the Sruhanavarnis Valley.

Sheared Dykes:- despite the overall low abundance of microgranite dykes there are a few dykes which have been sheared . In figure 4:29 the banding has been displaced by a deformed microgranite vein. The deflection of banding and the sigmoidal foliation within the microgranite shows sinistral offsets. The orientation of the vein is approximately 100-280° with dykes of this orientation more commonly having dextral offsets rather than sinistral offsets.

4:3:5:1 Thin-section

Microscopic evidence of deformation is generally very weak within these granitoids. The behaviour of the main mineral phases will be discussed separately.

Plagioclase:- essentially undeformed apart from very minor kinking. Only very faint PFC fabrics were observed in the SRU 2 tonalites with this alignment parallel to the foliation as indicated by the preferred orientation of biotite.

K-feldspar:- in the form of microcline, although micro-perthitic orthoclase is present. Within both plagioclase and K-feldspar there were no examples of any

Figure 4:28:- (a) Granite veins within pelite showing varying degree of boudinage depending their thicknesses. Sinistral and conjugate shear bands developed in neck of boudins.

Figure 4:28:- (b) Granite veins within quartzite raft-trains. This implies deformation was still occurring when host granite was more competent than the countryrock rafts.

Figure 4:28:- (c) Close up of boudinaged granite vein. Note the sinistral shear bands within the necks of the boudins.

fractured grains or any deformation related grain-size reduction. In the SRU 3 monzogranites the large K-feldspar megacrysts have been extensively myrmekitised.

Quartz:- this shows varying degrees of undulose extinction. The presence of deformation lamellae and partial subgrain development suggests minor recovery by dislocation climb processes has occurred (Passchier & Trouw 1996). Very limited dynamic recrystallisation has occurred as indicated by irregular grain boundaries suggestive of GBM recrystallisation. There is no obvious flattening within this mineral phase.

Biotite:- this is generally undeformed and showing weak alignment producing the S_6 foliation within these granitoids.

Muscovite:- this is seen as a sericitic alteration of plagioclase or lies parallel to the foliation in similar geometric relationships to biotite.

Figure 29:- E-W trending microgranitoid dyke which has sinistrally offset NE-SW trending granitic banding, NW Sruhanavarnis (offset 42cm).

4:3:6 The chronology of emplacement in the Sruhanavarnis valley

The spatial distribution of the various granitoid facies within the Sruhanavarnis Valley is summarised in figure 4:30.

The oldest granitoid phase in the Sruhanavarnis valley, based on field observations and detailed mapping, appears to be the fine to medium-grained SRU 1 granodiorite. Despite its presence within the Glenveagh 3A raft-zone it generally appears to be more abundant on either side of it, although the poor exposure to the SE of this raft-zone makes this less certain. To the immediate NW of this raft-zone

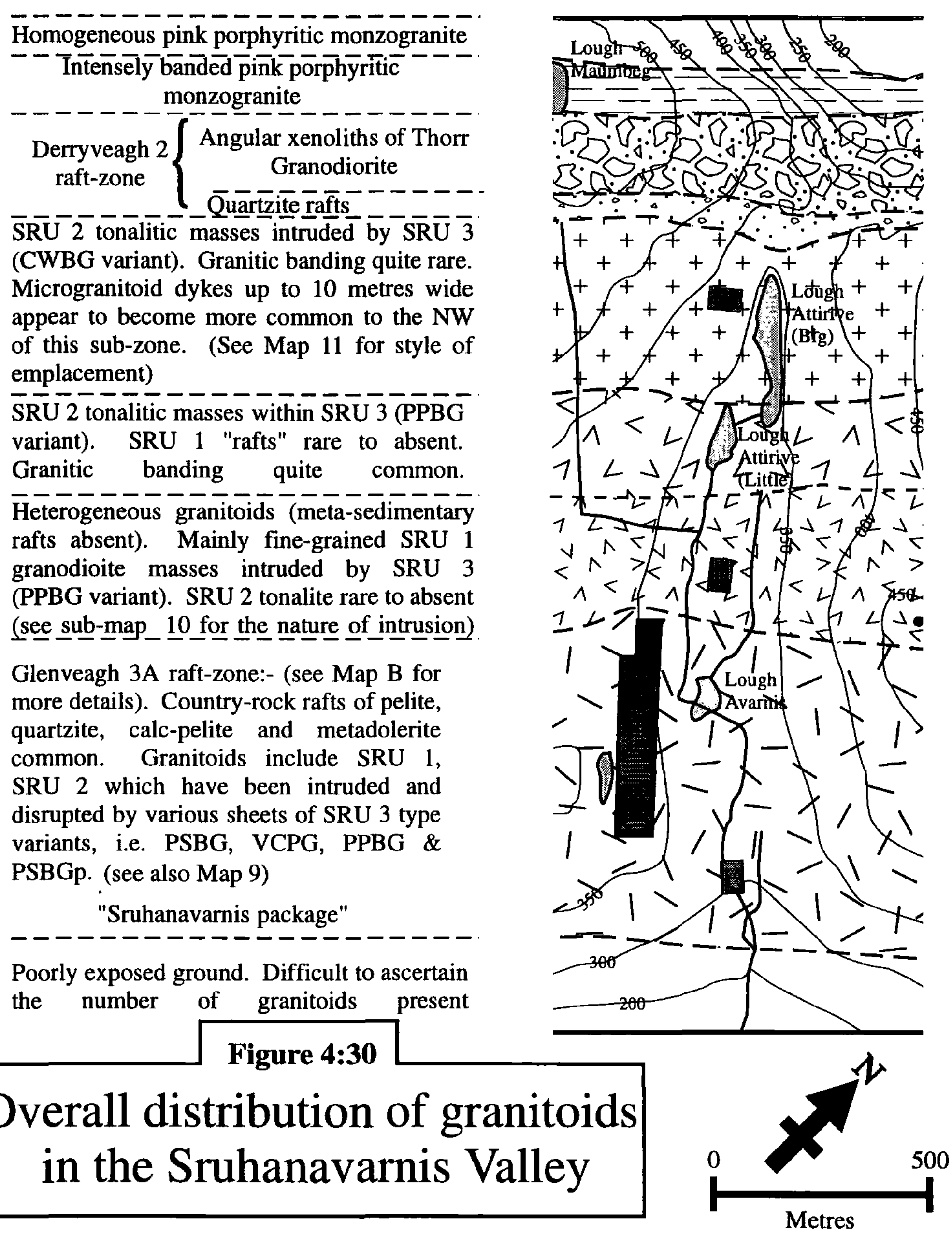


Figure 4:30

Overall distribution of granitoids in the Sruhanavarnis Valley

there is abundant SRU 1 type granodiorite. When all the subsequent granitoids are "removed" it can be envisaged that this was the early granite intruding into the Dalradian meta-sediments as possible sheet-like masses. This scenario would create the basic relationship observed with SRU 1 lying on either side of the countryrock raft-zone. This early phase of granodiorite is now preserved only as disrupted xenolithic masses. It is the authors belief that SRU 1 may have been fairly competent during the intrusion of subsequent granitoid phases as occasional angular xenoliths are found in younger granites. Furthermore some pegmatite veins within the SRU 1 granite have been truncated by later porphyritic monzogranites.

The next major phase of intrusion is the SRU 2 equigranular tonalites. Within the Sruhanavarnis area SRU 2 is more abundant than the older SRU 1 granodiorite, although its field relationships are essentially similar i.e. it occurs as rafts within younger porphyritic monzogranites. This tonalite is common *within* the Glenveagh 3A raft-zone and *between* the two major raft-zones in the NW part of the valley, although to the immediate north of the zone (for a distance of 250-300 metres) it has lower abundance. As with SRU 1 it appears that there were relatively high viscosity contrasts between SRU 2 and the SRU 3 monzogranites as indicated by quite sharp and planar contacts. The preferential alignment of these two older granitoids (usually parallel to the foliation) implies there may have been a structural control on fracture orientations during the intrusion of the younger monzogranites.

The most voluminous granitoid phase within the Sruhanavarnis Valley are the SRU 3 porphyritic monzogranites which are common within and between the Glenveagh raft-zones. Five variants have been observed in this area based on textural and modal differences in hand specimen. The author must emphasise that there was no unequivocal evidence to indicate the age relationships between these variants since inclusions of the different SRU 3 monzogranites within one another were not seen. The following suggestions about possible age relationships between the SRU 3 varieties are based on outcrop patterns in Map B. The VCPG monzogranite is in the form of a narrow sheet in sub-maps 1, 2 & 9. The presence of PSBG on either side might imply that VCPG is younger. In sub-map 7 there are masses of the PSBGp granite within PPBG possibly suggesting the former is older, a relationship which is reinforced in the NE square of sub-map 8. Age relationships between PSBG and PPBG were indeterminable. It is the authors belief that the SRU 3 granites are sheet-like pulses which have intruded and disrupted earlier formed granites. The finer-grained monzogranites appear to be earlier whilst the coarser varieties possibly younger (a feature also seen in the Glendowan Mountains; see section 4:5). The absence of distinct contacts between these monzogranites implies these porphyritic granites were emplaced within a relatively short time interval with partial coalescence along presently preserved transitional contacts.

The granitoids and country-rock rafts described up to now are intruded by a retinue of minor bodies. The earliest phase are the fine-grained, mafic-rich microtonalite dykes of which the best example is the discordant dyke in sub-map 9 (the "early transverse dyke" (figure 4:19) of Pitcher & Read 1960). This dyke is oriented at 120° and is interpreted by these authors as being intruded along embryonic joint systems within the cooling Main Donegal Granite. In sub-map 9 (figure 4:17) the age relationship of this dyke to the major intrusive phases in the Sruhanavarnis valley can be observed. The host granitoid along the majority of the length of the exposed dyke is SRU 3 (VCPG), although where it crosses the pelite raft (A2) it then intrudes through the SRU 2 tonalite. The margins of this dyke are commonly scalloped (Pitcher & Read 1960). The dyke is younger than the diffuse masses of pegmatite which occur in close vicinity to the raft-trains, whilst in regards to the later more regular pegmatite veins the dyke is older as it is often truncated, and sometimes displaced, by pegmatite. Although not shown on the figure 4:19 this early microgranite dyke has been intruded and truncated by later microgranitic-microgranodioritic dykes (see figure 4:17) which tend to have fairly regular widths and show little or no signs of disruption by the host granitoids. These "later" microgranite dykes are common within Map B and are generally parallel or sub-parallel to the foliation (056°) apart from one dyke visible on the SW side of Map 9 which is almost perpendicular to the foliation (133-313°). The intrusion of the pegmatites occur throughout the intrusion span of the microgranite dykes as there are localities where the microgranite truncate pegmatites whilst at other locations pegmatites truncate the youngest microgranites.

4:3:7 Discussion on the major granitoid phases

Within Map B the uncommon occurrence of composite rafts (e.g. country-rock, SRU 1 and SRU 2 etc.) created difficulty in establishing age relationships between the respective granitoids. For example the SRU 1 granodiorite is seen in contact with country-rock rafts but so are all the other types of granitoids. There are rare situations where the SRU 1 and SRU 2 granitoids are seen in contact and furthermore different lithologies of the Dalradian rafts are rarely seen together. Why is this so? The most likely explanation for this is that intruding granitic material will repeatedly inject along pre-existing contacts i.e. they exploit pre-existing viscosity\strength contrasts. Therefore one can envisage different host rock types behaving differently to applied stress (tectonic or magmatic) with this disharmony concentrating along contacts. An analogy of this was demonstrated when mapping this area where attempts to obtain fresh surfaces of granite contacts were made. In most cases the granites broke apart along such contacts when hit with a hammer. Therefore the most likely explanation is that the contacts of earlier granitoid phases tend be, but not always, exploited and intruded by later granitoids.

4:4 The spatial relationship of Crobane Hill and the Sruhanavarnis Valley

The Sruhanavarnis area and Crobane Hill lie approximately along strike from one another so the style of emplacement between these two areas will be compared and contrasted based on reconnaissance mapping of the granitoids in the intervening ground.

In figure 4:3 it was shown that the Crobane Hill mapping area lies between the Derryveagh 2 raft-zone and the Glenveagh 3A raft-zone. It was noted that these granitoids are generally heterogeneous with country-rock inclusions relatively uncommon. In the NW part of the Sruhanavarnis Valley the granitoids here, also lie between the two major raft-zones and are again quite heterogeneous with country-rock rafts also relatively uncommon. Therefore one might expect to observe some similarities between these two areas. On Crobane Hill the dominant granitoid phases were the porphyritic CBH 2 and biotite-poor, equigranular CBH 3 monzogranites which display an overall sheeted structure with the earlier CBH 1 granodiorite and possibly CBH 4* tonalites occurring as rafts within them. In the NW part of the Sruhanavarnis the dominant phases are two different types of porphyritic monzogranite (SRU 3: CWBG and PPBG) which contain raft-like inclusions of fine-grained granodiorite (SRU 1) and medium-grained biotite tonalite (SRU 2). Therefore the style of intrusion is similar in both areas but comparison of the hand specimens from the both areas reveals the granitoids are different in regards to their appearance. This implies the granites may not be laterally persistent over large distances, i.e. the original sheets are lensoid in nature. Bearing this in mind one might expect to observe granitoids reminiscent of both the Sruhanavarnis and Crobane Hill at locations in between these two areas. One such area is a well exposed section from the Bingorms northwards to the shores of Lough Slieve Snaght (see location maps 4:3 and 4:12). Finally, it has been observed that the granitoids to the SE of the Derryveagh 2 raft zone are much more heterogeneous than the porphyritic monzogranites to the NW of this. To test this hypothesis the granitoids were also looked at in the Kingarrow-Lough Veagh area.

4:4:1 Bingorms-Lough Slieve Snaght

The granitoids here are exposed on the western side of the hill, with has two unnamed 575m and 578m summits (GR B 9314), traceable northwards to the shores of Lough Slieve Snaght. At GR B 932142, immediately to the north of the pelite and metadolerite rafts belonging to the Glenveagh 3A raft-zone there is a monzogranite resembling CBH 3, although in this area its volume is considerably less than that seen at Crobane Hill. Within this monzogranite are the occasional rafts of the CBH 1 type granodiorite which are generally lensoidal in shape, aligned parallel to the weak foliation within these granitoids (058°). Walking northwards towards Lough Slieve Snaght the granitoids become more heterogeneous with rafts of the CBH 1 type granodiorite becoming more numerous, whilst the CBH 3 granite is longer visible. The host to these rafts are porphyritic monzogranites which partially resemble CBH 2 although they tend to be more variable in appearance when traced across strike (these

(these porphyritic granites vary due to slight colour changes in the K-feldspar megacrysts; size of the megacrysts and the grain-size variation of the groundmass). It appears that these changes may correspond to different sheets of monzogranite which slowly grade into one another with transitional boundaries separating them. Therefore there may be several different types of porphyritic monzogranite along this section, a feature which may not have been fully appreciated at Crobane Hill as the width across strike of the Map A was only 150 metres. In the Sruhanavarnis area though it was clear that the SRU 3 porphyritic monzogranites did vary across strike. From the Lough Slieve Snaght area it is apparent that the porphyritic monzogranites may consist of a series of compositionally similar sheets which have widths of 100-250 metres. In the absence of detailed mapping in this area, one cannot unequivocally prove this interpretation, although these thoughts were based on similar observations seen in the Sruhanavarnis valley. At GR B 932145 the coarse porphyritic host changes and becomes a coarse-grained porphyritic monzogranite host almost identical in appearance to the CWBG variant of the SRU 3 seen in the far NW of the Sruhanavarnis valley. Within this monzogranite rafts of medium-grained aphyric granodiorite-tonalite are seen. This style of intrusion continues up to the eastern shores of Lough Slieve Snaght where xenoliths of quartzite, pelite and then Thorr Granodiorite become more common, forming part of the distinctive Derryveagh 2 raft-zone. To the NW of this raft-zone, which is approximately 250 metres wide, the granitoids are very different in appearance and generally more homogeneous, a feature which has been observed at several other locations along this raft-zone and will be addressed in chapter 5 (section 5:3).

In the NW of the Sruhanavarnis Valley the granitoids are different from those seen on Crobane Hill, although the style of emplacement is the same. The CBH 2 is similar in appearance to the SRU 3 monzogranites, with it resembling the PPBG variant the most. The granodioritic rafts in the Sruhanavarnis are generally finer grained and contain less K-feldspar in comparison to their counterparts on Crobane Hill. Despite the granitoids of Sruhanavarnis Valley and Crobane Hill being different there are similarities between both areas as seen within the Lough Slieve Snaght area, i.e. the CBH 3 granite and the CWBG variant of SRU 3 are both observed along the section both of which contain rafts of earlier formed tonalite and granodiorite. The CBH 1 granodiorite is considerably coarser than the SRU 1 granodiorites with it showing greater resemblance to the SRU 2 tonalites in both appearance and from field relationships. For this reason it is believed that the CBH 1 granodiorite may have been penecontemperaneous with the intrusion of the SRU 2 tonalites.

4:4:2 Kingarrow-Lough Veagh

This area was looked at in a reconnaissance style with the main aim of studying the granitoids to the SE of the Derryveagh 2 raft-zone in the poorly exposed ground to the NW of Lough Veagh (see figure 5:24, p.234). From the work in other areas it is apparent that to the immediate SE of this prominent raft-zone heterogeneous granitoids are encountered, so it was the aim to see if such a relationship was true in this area.

The Derryveagh 2 raft-zone is encountered just to the south of Misty Lough South where it has a thickness of ~250 metres. Immediately to the south of the raft-zone there is abundant pegmatite within a relatively fine-grained host, although this pegmatitic component does become less abundant in a south-eastwards direction. Despite the poor exposure within this area the granitoids do show quite strong variation with medium-grained, equigranular granitoid and subtly porphyritic monzogranites present. These granitoids were homogeneous on scales of 50-100 metres, beyond which different looking varieties are observed. In some areas there are planar breaks of slope, (2 metres high) which represent granite contacts and can be traced along slope for up to 200-300 metres. At one example, GR C 019222, a weakly porphyritic monzogranite is in contact with a more resistant coarser porphyritic granite. The orientation of some of these contacts is between 045-060° approximately parallel to raft-trains of pelite and meta-dolerite which are seen just to the north of Lough Veagh and belong to the Glenveagh 3A raft-zone. Looking south-westwards from the above grid reference towards the hill called Keamnacally one can pick out a sheeted appearance to the hill with the orientation of the sheets being steeply inclined towards the south-east (see figure 4:31). Despite the poor exposure the granitoids of this area are quite heterogeneous and show some similarity to the NW Sruhanavarnis Valley and Crobane Hill. For this reason the granitoids which lie between the Glenveagh 3A raft-zone (also-called the "Sruhanavarnis package") and the Derryveagh 2 raft-zone will be called the "Crobane Hill-Lough Veagh package". This package is characterised by sheet-like masses of porphyritic monzogranite in which there are abundant rafts of older granitoid material and only minor amounts of countryrock metasediments and older plutons. The significance of this inter-raft-zone package will be addressed in chapter 8 of this thesis.

To the NE of this area the trend of the Derryveagh 2 raft-zone swings in a more southerly direction, by en echelon right stepping of the individual rafts (as described in Figure 10 of Pitcher & Read 1959). This change in trend causes the Derryveagh 2 and Glenveagh 3A raft-zones to converge and greatly narrows the Crobane Hill-Lough Veagh package.

In the following section the granitoids to the south of the Glenveagh 3A raft-zone will be discussed. The lack of exposure within the fault controlled Glenveagh and Owenbarra valleys means the granitoids to the immediate south of this raft-zone cannot be studied.

Figure 4:31:- Photograph looking SW toards Keemnacally with this mountain appearing to be composed of granitic sheets inclined steeply to the SE.

4:5 Glendowan Mountains

Figure 4:32:- Photograph of the Glendowan Mountains taken from Crockbrack (near Sruhanavarnis). Moylenanav in the midground with Crockskallabagh behind it on the left, with Crockastoller further away to the right.

Introduction

The Glendowan Mountains are situated to the south of the Owenbarra river (due south of the Sruhanavarnis Valley) within the central regions of the Main Donegal Granite. The main hills within these mountains are Moylenanav (539m), Crockskallabagh (494m), Croaghacullin (435m) and Crockastoller (418m). The granitoids in this area were studied on a traverse from Croaghacullin to Moylenanav. Despite good exposure in this area it is rarely as continuous as that seen within the Derryveagh Mountains. The main point of access to this area is from the R254 just to the NE of Lough Barra, along the south-western flanks of Moylenanav. Access to the Croaghacullin side is best obtained from the R254 in the Bullaba valley (near the confluence of Croaglughoge Stream with the Bullaba River (GR C 983149)).

The granitoids within this area are quite heterogeneous with six facies having been identified based on morphological and petrographic differences. As seen at Crobane Hill and the Sruhanavarnis Valley the compositions range from aphyric granodiorites and tonalites to porphyritic monzogranites (see figure 4:33). Country-rock material is generally low in abundance apart from two pelitic raft (+ metadolerite) swarms on the SW flanks of Moylenanav and a similar swarm on Crockastoller. The former of these swarms belongs to possibly the Binaniller 3B raft-zone whilst the latter swarm may be part of the Crockmore 4 raft-zone (Pitcher & Read 1959).

In this area two detailed maps, on a scale of 1:250, were completed, one of which was made on the well exposed ridge 1 km due south of the summit of Moylenanav (GR B 955125) whilst the other was undertaken within a well exposed eroded peat bog 0.75 km to the SE of the summit of Croaghacullin (GR C 982122). The size of the maps 60 × 90 metres. The absence of continuous and good quality exposure is the reason for the small size of these mapping areas. The granitoids of the Glendowan Mountains will be addressed in two sections and will be prefixed by "GM" (i.e. Glendowan Mountains).

4:5:1 Croaghacullin

The main areas of exposure here are the elongated NE-SW ridge of Croaghacullin and also an area to the SE of this summit in heavily eroded peat bogs in which the granitoids are excellently exposed. Map I (figure 4:34) was undertaken in this latter area (GR B 982122). The granitoids are quite heterogeneous in this area with five main granitic facies identified, ranging from granodiorites to tonalites through to monzogranites. These are all identified in Map I and it is from this map the granitoids will be described in regards to their general petrography and field relationships.

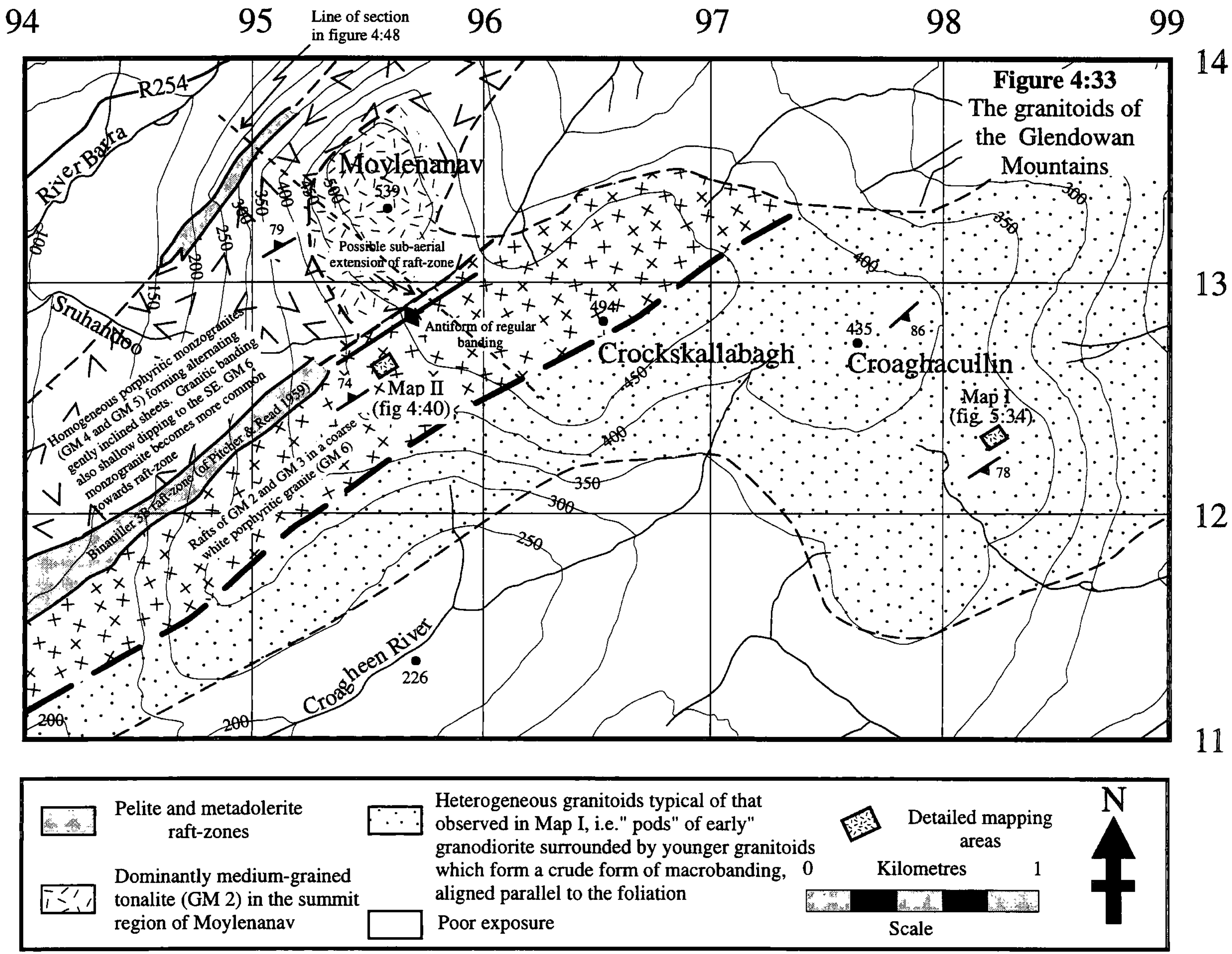


Figure 4:33 The granitoids of the Glendowan Mountains

4:5:1:1 The granitoid facies of Croaghacullin

Map I (figure 4:34) shows a crude form of compositional banding to the granitoids in this area with these phases being discussed in approximate order of age based on relationships seen in this map, and from other adjacent areas. Photographs of polished hand specimens of the "GM" granitoids are shown in Appendix A.

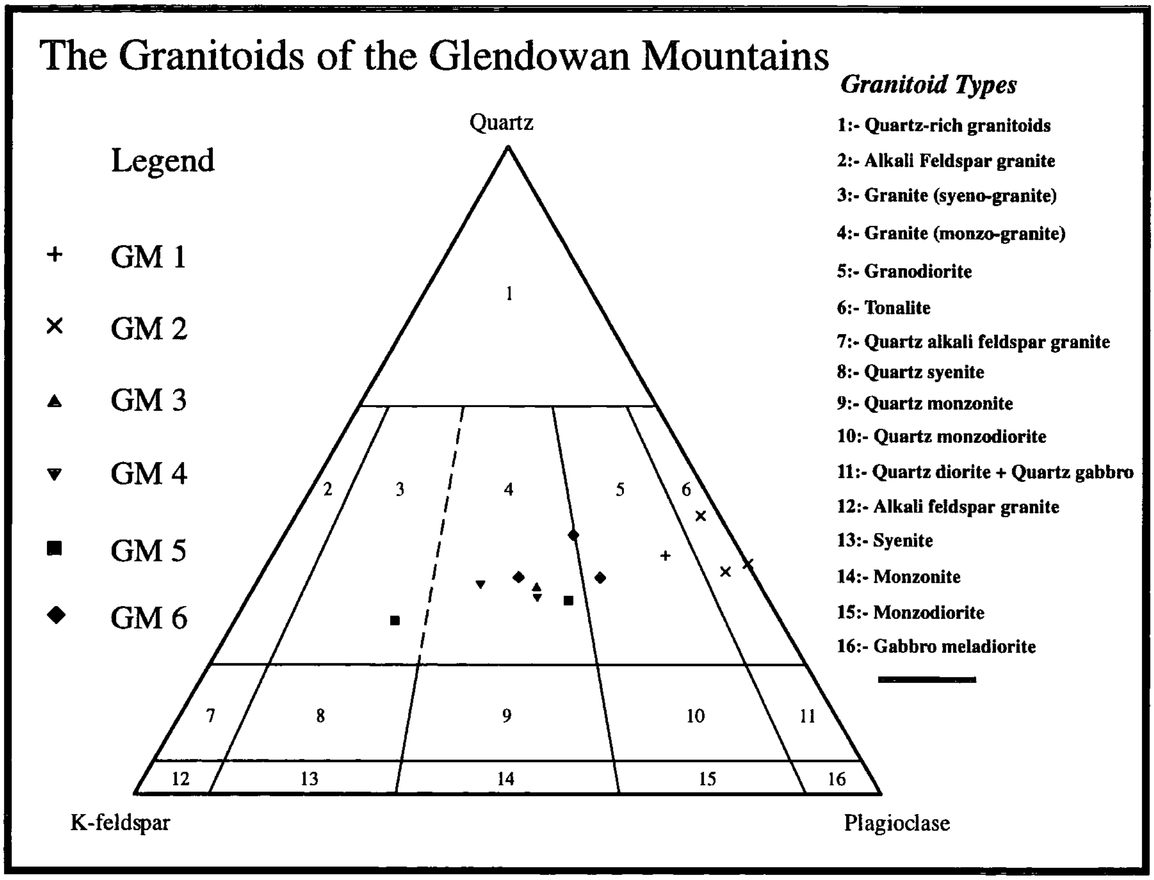


Figure 4:35-: QAP plot for the granitoids of the Glendowan Mountains

GM 1

This is the finest grained granitoid within the mapping area with grain sizes less than 2 mm and its modal proportions plotting in the granodiorite field of the QAP plot. This granodiorite is composed of white plagioclase (45%), grey quartz (31.5%), olive-green biotite (13%) K-feldspar (10%), epidote (0.3% and other accessories (0.2%). In hand specimen this granodiorite has a dark-grey appearance (due to high biotite content), whilst on weathered surfaces the general absence of coarse feldspar also gives this granitoid a very dark-grey colour allowing it to be clearly distinguished from the other varieties in this area. Within Map I the GM 1 granodiorite occurs as a relatively homogeneous mass within the central areas of the map. Along the margins of this mass there are smaller xenoliths of smeared out GM 1 within younger granitoids.

GM 2

This equigranular tonalite has a grain size between 2-3 mm, although there are larger, euhedral plagioclase phenocrysts with lengths between 2-4 mm. The main components are white plagioclase (52.5%), grey quartz (33%), olive-green biotite (11%) K-feldspar (2%), muscovite (1.25%) and epidote (0.25%). Other accessories include zircon and apatite. The colour of this tonalite is medium to dark-grey tending to be slightly lighter in colour, when compared to GM 1, due to the larger grain-size of the feldspars. Volumetrically GM 2 is not in high abundance in the Croaghacullin area. In appearance on fresh surfaces, this tonalite shows strong similarities to the SRU 2 tonalites seen in the Sruhanavarnis Valley. Within Map I it outcrops as narrow bands which are broadly parallel to the foliation within the pluton and smaller fragments of this tonalite are seen in all granitoid facies apart from GM 1.

GM 3

In contact with the GM 1 granodiorite is a medium-grained, aphyric monzogranite (grain size 1-3 mm) which is composed of white plagioclase (35%), white microcline (28%), grey quartz (29.5%), olive-green biotite (7%) and muscovite (0.5%). The lower biotite content within GM 3, when compared to the GM 2 tonalite, gives this monzogranite a much lighter grey colour in both fresh and weathered hand specimens. Its grain size is intermediate between the GM 1 and GM 2 granitoids and hence it is sometimes very difficult to distinguish from weathered surfaces alone. Thin-section reveals the considerably higher K-feldspar content within this monzogranite when compared to GM 1 and GM 2.

GM 4

The GM 4 monzogranite has a tendency to be porphyritic with the 3-5 mm inclusion-rich microclines sometimes exceeding the grain size of the groundmass which is 2-3 mm. It is composed of white plagioclase (32.5%), white K-feldspar (31.5%), grey quartz (29.5%), olive-green biotite (5.3%), muscovite (1%) and other accessories (0.2%) The appearance is very similar to that of the GM 2 tonalite due to the subtly porphyritic nature of this monzogranite. Weathered surfaces are lightish-grey and often it is only distinguishable from GM 2 during good sunlight. It contains streaked-out inclusions of the earlier GM 1 and GM 2 granitoids which are all oriented parallel to the foliation (i.e. 058°).

GM 5

The GM 5 facies is the coarsest granitoid in the Croaghacullin area having a grain size of 3-4 mm with megacrysts of microcline up to 8 mm in size. The main components are white plagioclase (30%), pink K-feldspar (36%), grey quartz (28.5%), olive-green biotite (3%), muscovite (2.5%). A sample of this monzogranite was found to be the most alkalic granitoid encountered within the pluton plotting in the syenogranite field on the Streikeisen plot. In both weathered and fresh surfaces the coarse nature and the pink colour of the K-feldspar megacrysts allows it to be identified easily.

4:5:1:2 Field and age relationships in the Croaghacullin area

The exact age relationships between all the granitoids on Croaghacullin is uncertain but from the relationships observed the following conclusions can be made:-

1) The GM 1 granodiorite is the oldest as xenoliths of it are all found in the other facies present.

2) Study of figure 4:36 shows several key relationships between the GM 2 tonalite (darkish granitoid on the left) where it is intruded by bands of lighter GM 3 monzogranite. These bands are then truncated by the GM 4 porphyritic monzogranite. The planar and sharp nature of this latter contact implies that both GM 2 and GM 3 (and by inference the GM 1 granodiorite) may have been highly competent during the later intrusion of GM 4 and by inference the GM 5 monzogranites.

3) The age relationship between these latter porphyritic monzogranites, (i.e. GM 4 and GM 5) is uncertain due to the absence of xenoliths of the one in the other and truncation features i.e. if one monzogranite truncated a pegmatite vein within the other. The presence of the GM 4 band in GM 5 on the right of Map I may suggest that the consistent width is due to it being a dyke, although one must emphasise that this evidence is a little equivocal. The contacts between these two monzogranites are very subtle and are commonly only clearly visible when viewed from a distance of 2-3 metres. The majority of the contacts within this mapping area are oriented parallel to sub-parallel to the foliation and this produces a crude form of compositional macrobanding.

The style of emplacement in Map I was compared with the granitoids on the summit of Croaghacullin. From the five granitoid facies already documented only three were encountered on this summit. These were GM 1, GM 3 and a much greater relative volume of GM 4. The GM 1 granodiorite occurs as large "pods" within the other two granitoid types, often having dimensions of up to 10 × 30 metres with the

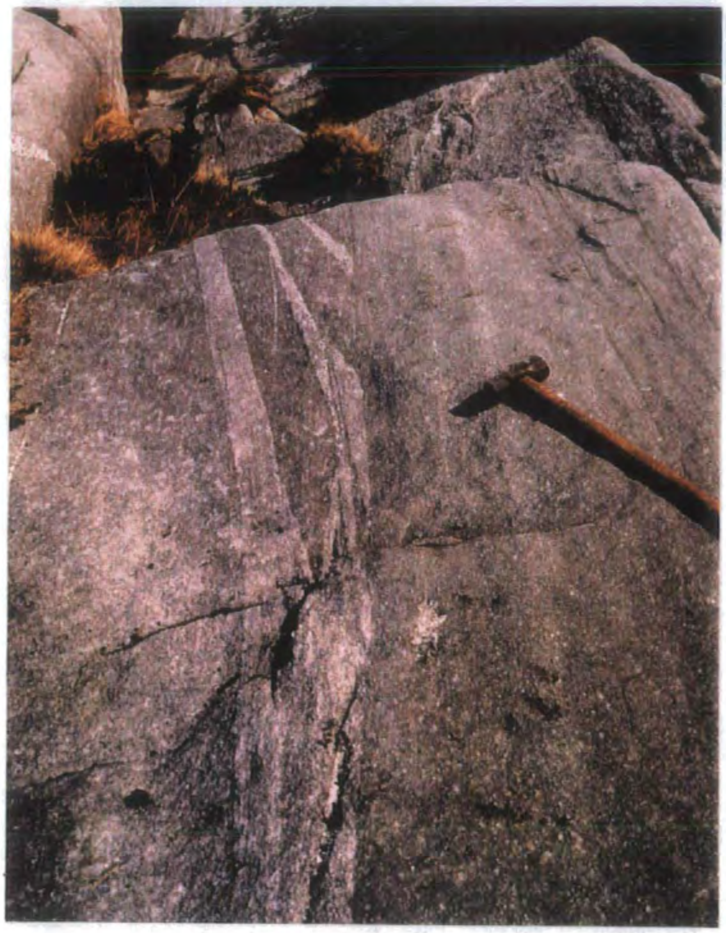


Figure 4:36
Bands of GM 3 within GM 2 (left). Both these facies have been truncated by the GM 4 porphyritic monzogranite.

Figure 4:37:- Complex banding developed around the GM 1 mass in Map I. GM 1 forming the homogeneous granitoid on the left with bands of GM 2-4 on the right surrounding it.

long axes of such pods aligned parallel to the foliation. As many as ten of these large pods were observed on the elongated summit of Croaghacullin. A common occurrence around these pods is a type of granitic banding which tends to quite irregular in nature. Figure 4:37 shows an excellent example of this type of banding which develops around the GM 1 pods. In some areas on the photo the banding is discordant i.e. one set of banding is truncated by a second set. Pitcher & Read (1959) observed a similar phenomena on the summit of Croaghnagapple, located 1.5 km to the ENE of the Map I area.

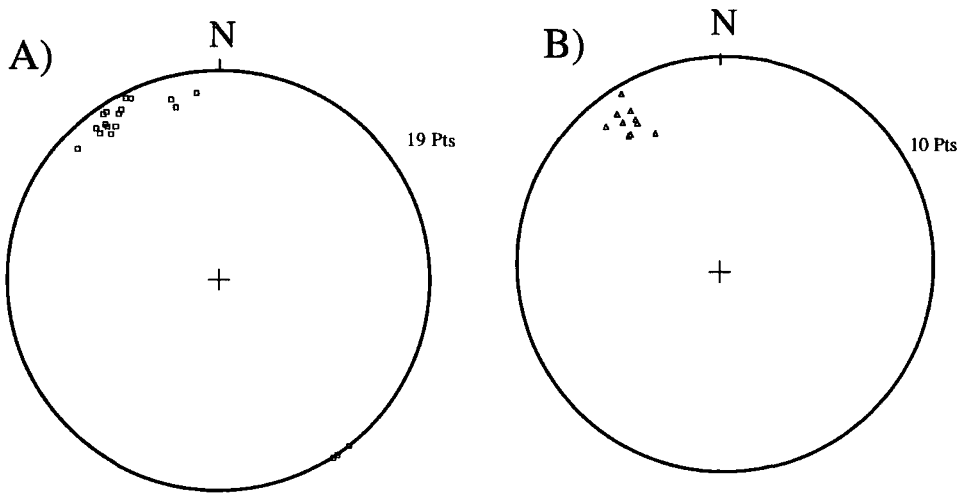
4:5:1:3 Deformation

The foliation within these granitoids is generally stronger than compared with the Sruhanavarnis area and Crobane Hill (see figure 4:38). Features such as banding also suggest some syn-plutonic deformation has occurred with it tending to be parallel to the S_6 foliation . The deformation in this area will be dealt with at both the field and thin-section scale.

4:5:1:3a Outcrop

The presence of banding around all of the GM 1 granodiorite pods implies this granitoid was competent during deformation whilst the later granitoids, (see figure 4:37) GM 2 and GM 3 appear to have been less viscous and have been smeared out around it. No shear sense can be seen in the bands themselves although they are cut by late stage small-scale faults or shear bands (see figure 4:37) oriented 022-202° which show dominantly sinistral offsets. In Map I the foliation and smearing out of xenoliths (including Thorr Granodiorite) is also most intense around the GM 1 masses implying high strains were localised around the margins of these pods.

Within the map area small-scale ductile (outcrop scale) dextral shear zones can be observed. In square B2 displacements, marked by the offset of GM 1 xenoliths, of two metres are visible. The orientation of this shear zone is 098-278° and corresponds to an anti-reidel direction. Within the marginal areas of the pluton as a whole the orientation of antithetic C' foliation is very similar to the orientation of these small shear zones present in the more central regions of the pluton implying these latter regions of the pluton were being subjected to the same deformation. Generally the main foliation within this area is the S_6 foliation with no S-C' fabrics developed apart from a 10 metre wide, NE-SW striking zone at the summit of Croaghacullin. The sense of shear within this small shear zone is sinistral.



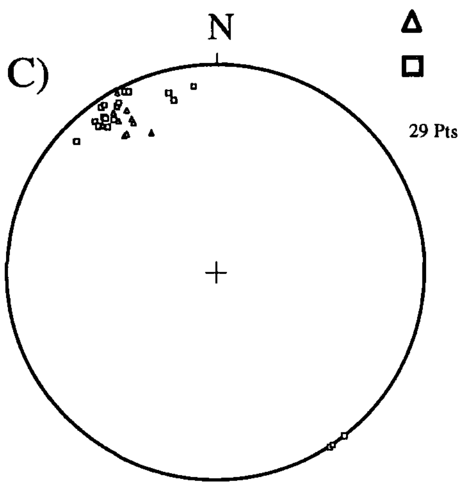


Figure 4:38

Croaghacullin area, Glendowan Mountains

A) S_6 foliation

B) Granite Banding

C) Composite plot of both S_6 and banding

4:5:1:3b Thin-section

In thin section the granitoids are foliated to varying degrees and the preferred alignment is most noticeably produced by biotite.

PFC fabrics:- within the GM 1 granodiorite faint PFC fabrics were seen marked by preferred alignment of euhedral to anhedral plagioclase (figure 4:39a) together possible evidence for magmatic tiling with the sense of rotation indicating sinistral shear. Within this granodiorite CPS overprinting is relatively minor with quartz only showing slight evidence of intracrystalline deformation. PFC and stronger CPS fabrics were observed in the GM 3 monzogranite where the alignment of plagioclase laths is generally parallel to the foliation, the latter depicted by the alignment and strong flattening of quartz. Brittle microfracturing has occurred in some plagioclases with the cracks having been infilled with quartz which has behaved in a ductile fashion (figure 4:39b). Also in the same slide many of the plagioclases showing sweeping undulose extinction due to sub-microscopic brittle kinking. Bouchez *et al.* (1992)

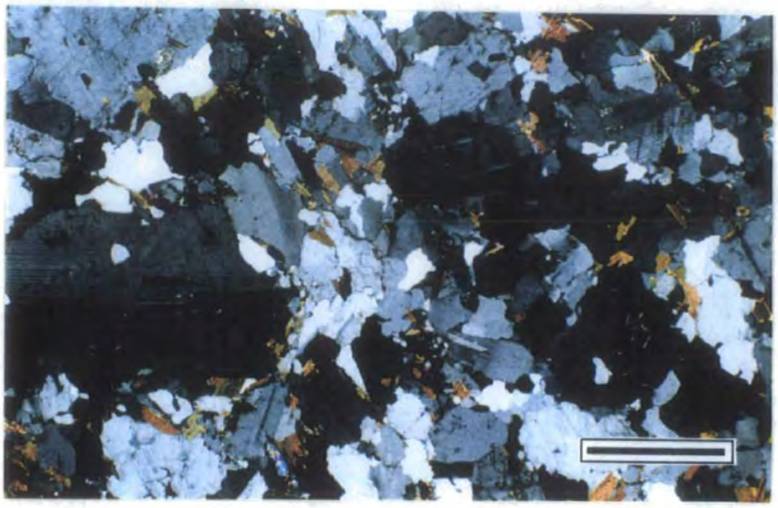


Figure 4:39 :- (a) Weak PFC fabrics within the GM 1 granodiorite (trending E-W) (scale bar =1mm).

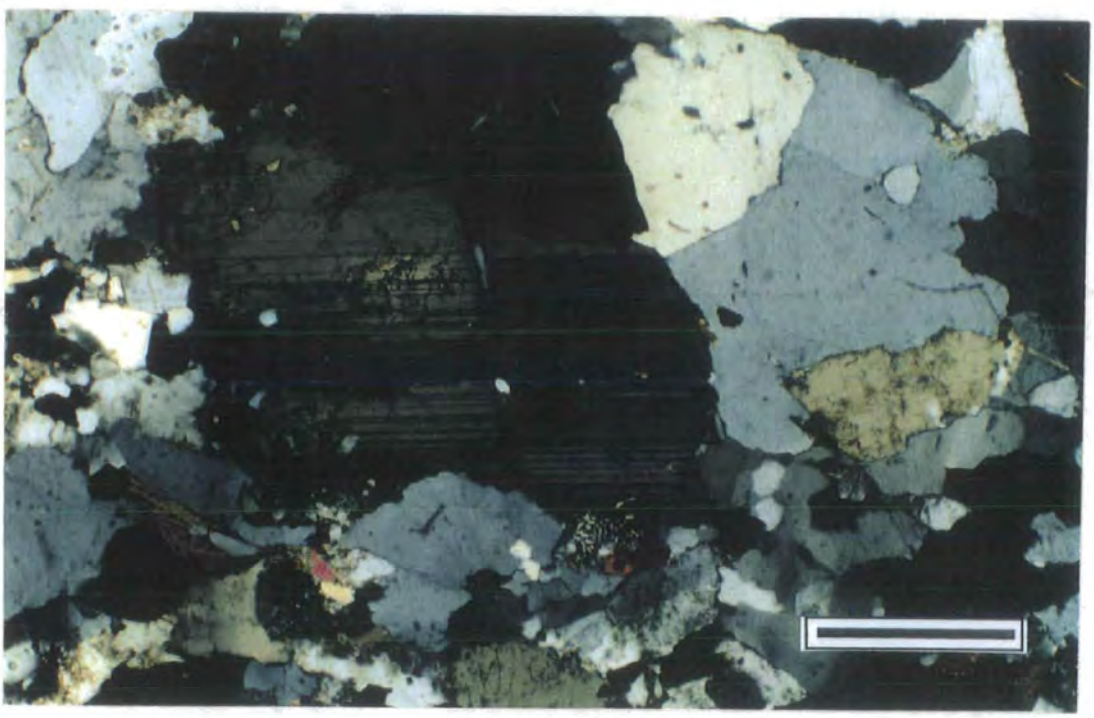


Figure 4:39:(b) Sub-magmatic fracturing of plagioclase. The presence of weakly strained quartz nearby argues against solid-state fracturing. (GM 3 monzogranite) (scale bar = 1mm).

have stated that such features can form in the sub-magmatic state, (i.e. below the RCMP but still with interstitial melt.). At this stage of crystallisation the early formed phases such as plagioclase form a rigid framework which can transmit stress. Bouchez *et al.* (1992) stated that fracturing may occur where grains are in contact with one another. These fractures will then become filled with melts composed of quartz, albitic plagioclase or K-feldspar (low melting temperature mineral phases). Bouchez *et al.* (1992) stated that within thin-sections to be studied it was important that quartz had not undergone any recrystallisation as evidence of this may imply the feldspars may have been fractured during solid-state deformation. Within all the granitoids studied quartz shows strong undulose extinction, although there is only limited recrystallisation (SR). The presence of lobate and irregular margins between quartz grains suggests GBM recrystallisation. The deformation structures now preserved within quartz and the presence of myrmekite in microcline implies relatively low-temperature deformation, i.e. ~ lower-mid greenschist facies.

In conclusion the occurrence of both banding (interpreted as due to deformation) and PFC fabrics which are sub-parallel to the foliation implies that there has been deformation in the magmatic state, sub-magmatic state (below the RCMP) and during the solid-state.

4:5:2 Moylenanav - Crockskallabagh

The exposure between Croaghacullin and Crockskallabagh is very poor so it is difficult to trace the geology across this area. Some of the granitoids already described in Map I are observed in this area as well as some new varieties. The granitoids in this area are generally well exposed on the NW and SW flanks of Moylenanav whilst on the NE side of this mountain the amount of exposure is very poor due to a thick covering of peat. The ground 0.75 km to the south of the summit of Moylenanav is also well exposed forming a prominent NE-SW trending ridge. It was on this ridge where Map II (figure 4:40) was undertaken. Between the summits of Moylenanav and Crockskallabagh exposure is moderate with large isolated masses of granitoid commonly surrounded by thick peat. On the NW side of the latter summit the exposure is generally excellent composed dominantly of homogeneous porphyritic monzogranite.

4:5:2:1 The granitoid facies of the Moylenanav-Crockskallabagh area

This area studied in Map II is located 100 metres to the SE of a prominent pelite raft-zone, which forms the supposed Binaniller 3B raft-zone of (Pitcher & Read 1959) as shown in figure 4:33. Apart from the presence of a few rafts of heavily digested pelite and locally derived leucosomes the main granitoid types are medium to fine grained granitoids which resemble the GM 2 and GM 3 granites described in the Croaghacullin area. These two granitoid facies behave as rafts within a coarse, white, porphyritic host.

GM 2

The appearance of this tonalite in hand specimen is identical to that described in Map I although in this area it is more abundant. The grain size of this tonalite (mafic content ~13%) is 2-3 mm, although plagioclase feldspar megacrysts are up to 5 mm in size. The biotites in GM 2 are red-brown in thin-section whilst those in GM 2 of Map I were olive-green in colour. In Map II this tonalite was only observed in contact with the GM 6 monzogranite.

GM 3

This monzogranite is identical to the GM 3 type already described in Map I. The grain size is 1-2 mm apart from the sparsely distributed K-feldspar megacrysts which are up to 5 mm in size.

GM 6

This is the most coarse-grained granitoid (monzogranite) seen in the Glendowan Mountains and has a groundmass of 2-4 mm, with microcline megacrysts up to 12 mm in diameter. The average modal percentages are white plagioclase (32.5%), white microcline (29%), grey quartz (32%), red-brown biotite (6%) and muscovite (0.5%). This porphyritic monzogranite is distinguished from the GM 5 monzogranite by the coarser grain size and the colour of the K-feldspar megacrysts, which in this granite are white in comparison to the pink microclines of GM 5. GM 6 is identical in appearance to the VCPG variant of the SRU 3 monzogranite seen in the Sruhanavarnis Valley. In Map II it is the dominant granitic phase having intruded into the earlier GM 2 and GM 3 granitoids. Along strike to the NE of this area, in between the summits of Moylenanav and Crockskallabagh this monzogranite is very common. Further to the NE, on the NW side of Crockskallabagh, GM 6 is unusually homogeneous with no inclusions of earlier granitoid phases encountered, despite the excellent exposure.

4:5:2:3: Field relationships in the Moylenanav area

Map II

Both the GM 2 and GM 3 granitoids tend to form rafts within the GM 6 monzogranite with their long axes parallel to the S_6 foliation. The behaviour of these early granitoids during the intrusion of GM 6 seems to be dominantly brittle with the contacts of these earlier phases tending to very sharp with biotite "wisps" separating them from the GM 6 monzogranite. Furthermore within these tonalitic and monzogranitic rafts there is a prominent set of joints (not present in GM 6) which appear to have controlled their shape during the intrusion of the GM 6 monzogranite (see Map II: figure 4:40). Two implications can be made from this observation.

i) The consistent orientation of the joints (114-294°) within the different GM 2 and GM 3 rafts implies there has been minimal rotation of these rafts during intrusion of GM 6, i.e. there was a preferential structural control during the intrusion of the GM 6 into the older GM 2 and GM 3 granitoids.

ii) The joints must have been present before the intrusion of the GM 6 monzogranite, i.e. the GM 2 (uncertain about GM 3) was highly competent, evidence reinforced by angular and sharp contacts controlled by the joints.

Due to this mapping area lying on a slope it was observed that rafts of GM 2 and GM 3 are often separated both vertically and horizontally by younger GM 6.

Moylenanav

Further to the NW of Map II on the SW flanks of Moylenanav the granitoids are quite homogeneous with two main varieties identified. The coarser pink variety is almost identical in appearance to the GM 5 monzogranite of Croaghacullin, (although in this area this pink variety appears to contain more muscovite) whilst the finer grained variety is similar to the weakly porphyritic GM 4 monzogranite. Figure 4:41 is a photograph of the SW flanks of Moylenanav (viewed from Lough Barra) where these two monzogranites possess an overall sheeted form. There appears to be three main sheets which dip at a shallow angle of 25-30° to the SE. The upper sheet is composed of very homogeneous GM 5 type monzogranite. Towards the base of the upper sheet there is banding which is oriented essentially parallel to the sheet contacts (i.e. ~ 035°\50°SE) where GM 5 slowly grades into finer-grained GM 4, over a distance of 20-30 metres by alternating bands of the two granites. Below this GM 4 sheet there appears to be another sheet of coarse pink porphyritic monzogranite resembling GM 5. The banding separating these gently inclined sheets strongly resembles the "regular" banding described by Pitcher & Read (1959) and Berger (1971) (see figure 4:42).

The contacts between these major sheets is commonly marked by gullies which may represent eroded out meta-sedimentary rocks. At GR B 948134 vertical sections of raft-trains composed of pelite and meta-dolerite are clearly visible and inclined moderately to the SE (see figure 4:43) thus supporting the idea that these raft-trains separate different granitic sheets. This group of raft-trains are laterally persistent and traceable along the NW flanks of Moylenanav for about 700 metres,

Figure 4:41:- View of the SW flanks of Moylenanav showing orientation of sheets of monzogranite gently inclined to the SE.

Figure 4:42:- Regular type banding developed along the boundaries of the major monzogranite sheets (in fig 4:41) and inclined gently to the SE (SW flanks of Moylenanav).

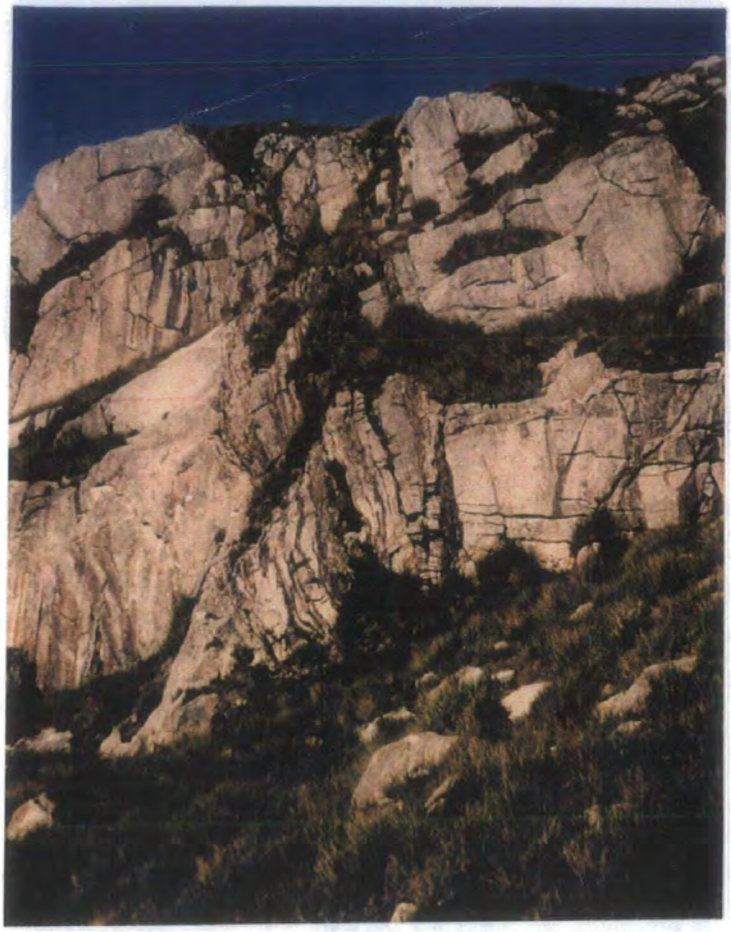


Figure 4:43
Vertical section through a pelite raft-train on the SW flanks of Moylenanav.

Figure 4:44:- on the SW flanks of Moylenanav where the sheeted nature of the porphyritic monzogranites is seen more clearly. Small gullyies may represent eroded out pelite, i.e. sheet boundaries.

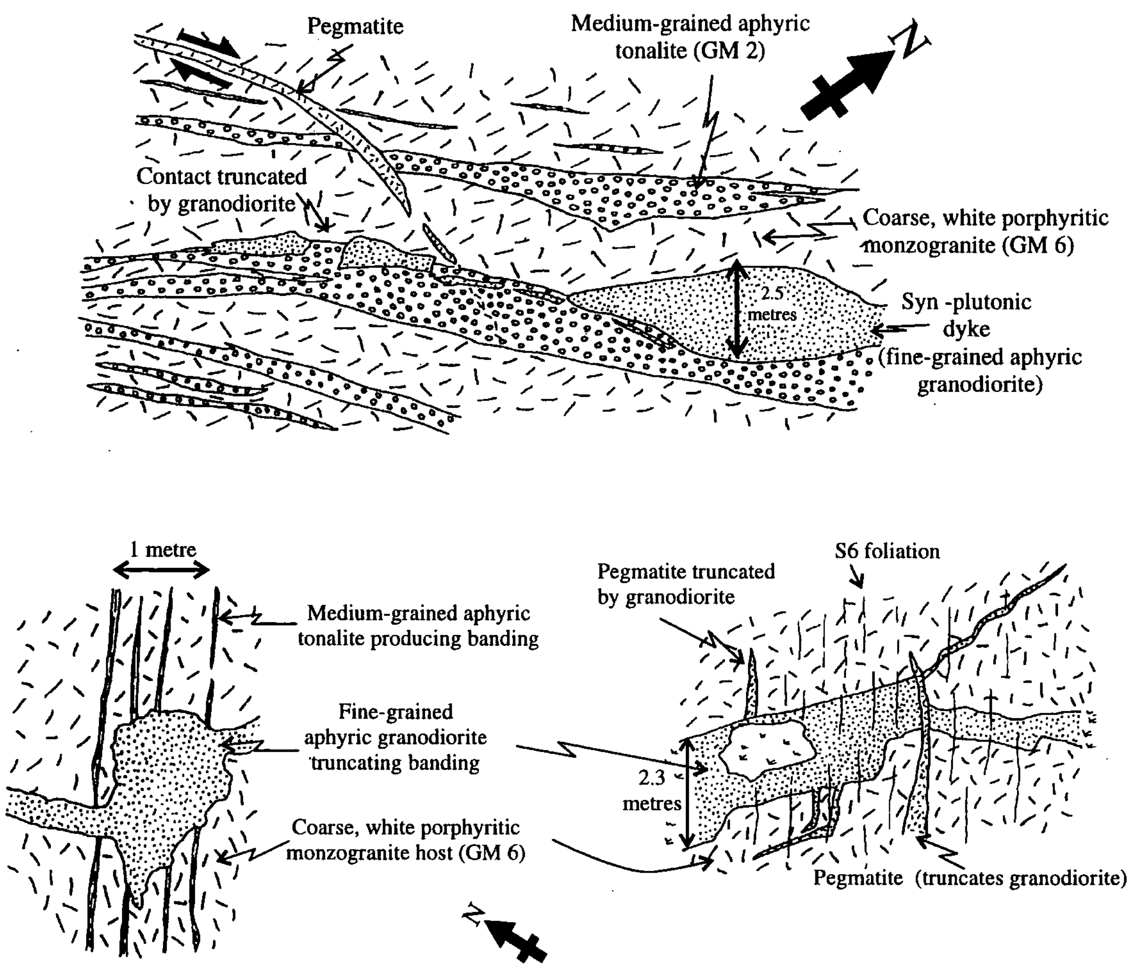
just above the vegetation line. In figure 4:44, there are a series of gullies on the south-western flanks of Moylenanav which may correspond to eroded out pelites. Further isolated raft-trains of pelite are found at GR B 945125 to the SW, and along strike, from the gullies on the SW flanks of Moylenanav suggesting they might well lie along the boundaries of the GM 4 and GM 5 sheets. Overall the granitoids to the NW of the Glenveagh 3B raft zone are very homogeneous with other granitoid facies, notably the fine-grained GM 1 granodiorite tending to be absent. Occasional masses of the GM 2 tonalite are seen and form a type of "xenolithic" banding. In the summit region of Moylenanav there is a large mass of unbanded tonalite resembling the GM 2 tonalite which has locally been intruded by the GM 6 porphyritic monzogranite. The relationships suggest that in the summit region the earlier formed GM 2 granite body has governed the orientation of later sheets and banding (i.e. syn-plutonic deformation) during the emplacement of the later granitic phases (i.e. GM 4, GM 5 and possibly GM 6).

The majority of the pelite rafts in this area belong to the Glenveagh 3B raft-zone (060-240°) which is up to 120 metres wide and is mainly composed of heavily assimilated pelite and to a lesser extent metadolerite. Within this zone pelite comprises up to 70% of the rock present. Apart from pegmatite, the dominant granitoid phase within the raft-zone is the GM 2 type tonalite. The boundaries of the raft-zone are relatively sharp and demarcated to the SE and NW by prominent gullies. At GR B 952123 the raft-zone terminates in outcrop. It is likely that this raft-zone continues below predominantly GM 6 porphyritic monzogranite. The presence of regular banding to the NE at topographically higher levels may suggest its presence at depth (see later).

Within the Moylenanav-Crockskallabagh no "pods" of the GM 1 granodiorite were observed, although there are granitoids identical in appearance to GM 1 in this area which intrude into all of the other facies. The field relationships in figure 4:45, from the SW flanks of Moylenanav suggest that this fine-grained granodiorite-tonalite may belong to a separate suite of syn-plutonic dykes that have been re-mobilised after emplacement.

The granitoids to the immediate SE of the Binaniller 3B raft-zone are generally typified by GM 2 and GM 3 rafts within the GM 6 porphyritic host, as shown in Map II. Approximately 200 metres further to the SE of Map II the GM 6 facies is absent and the granitoids and their style of emplacement here resembles that seen in the Croaghacullin area, i.e. pods of GM 1 surrounded by younger granitoids to produce a crude form of compositional banding.

A) GR B 947124:- disrupted fine-grained granodiorite dykes



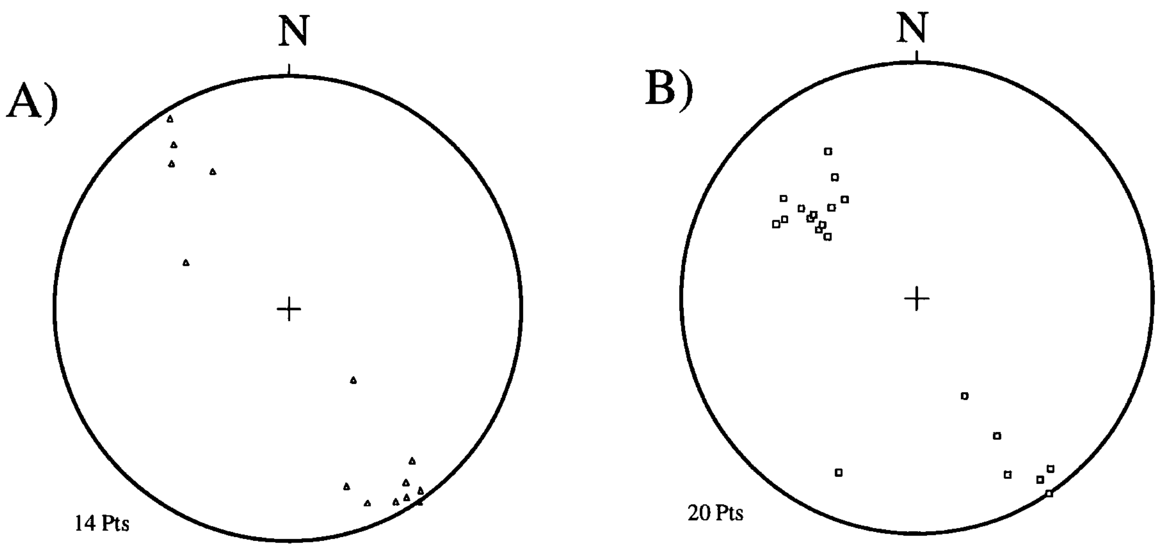
B) 40 metres from A):- irregular-shaped granodioritic mass truncating banding

C) same location as B):- overlap between intrusion of fine-grained granodiorite and pegmatites

Figure 4:45:- fine-grained granodiorites in the Moylenanav area which post-date all the porphyritic monzogranite phases. These intrusives are identical in appearance to GM 1.

4:5:2:4 Deformation

In comparison to the Croaghacullin area, the degree of deformation within this more central area of the pluton is generally much lower. In the thin-sections collected from the Moylenanav area no magmatic fabrics were observed and plagioclase laths had near random orientations. The foliation within these granitoids is only marked by very faint alignments of biotite. Quartz, although showing undulose extinction, displays no evidence of any flattening with dynamic recrystallisation within these granitoids at a minimum.



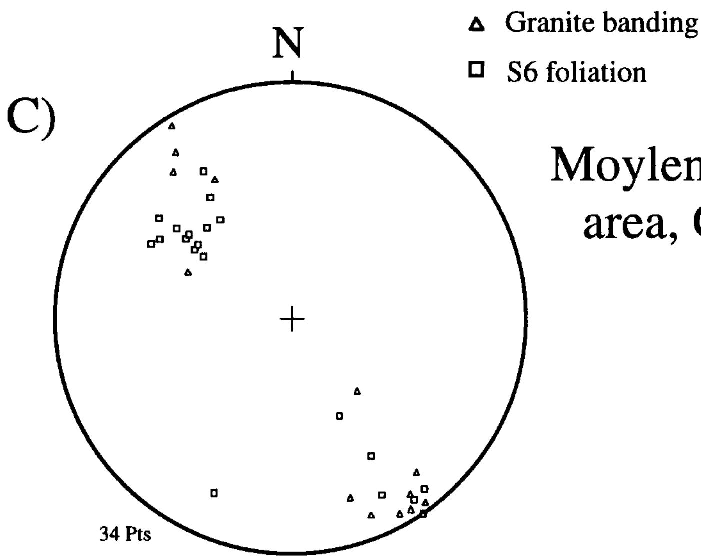


Figure 4:46

Moylenanav-Crockskallabagh area, Glendowan Mountains

A) S6 foliation

B) Granite Banding (regular type)

C) Composite plot of banding and S6 foliation

Outcrop

Granitic Banding:- the banding seen in the Moylenanav area is similar to the regular banding observed in the Doocharry area, with both tending to show a more discordance to the typical orientation (~ 058°) of the foliation in the majority of the pluton. Figure 4:46 a; b & c shows equal area stereoplots of banding, foliation and composite banding-foliation. In comparison to Croaghacullin both the banding and foliation are more variable in this area, but overall show a reasonable degree of similarity. On the NW and SW flanks of Moylenanav banding dips gently towards the SE. In the vicinity of the mapping area and the Binaniller 3B raft-zone the banding becomes more parallel to the foliation in the pluton i.e. 058° and is steeply inclined (see figure 4:47). It is in this area that the dip direction of the banding changes.

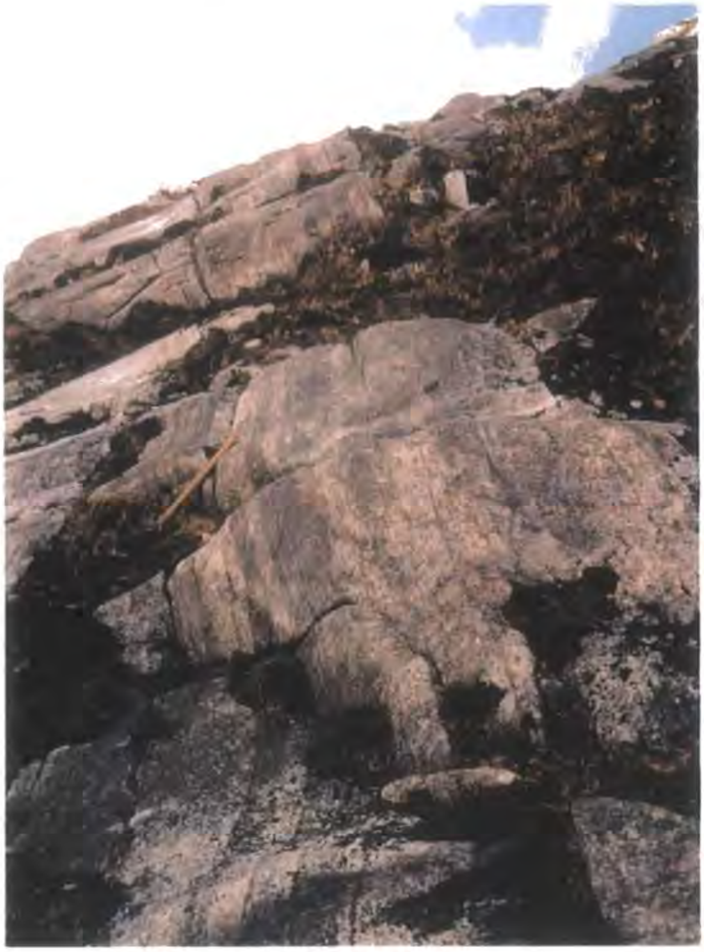


Figure 4:47
Regular banding steeply inclined to the SE. Situated to the immediate SE of the Binaniller 3B raft-zone and 300 metres to the NE, along strike from Map II (fig. 4:40).

In Map II the banding dips towards the SE whilst at GR B 959129 (to the SE of Moylenanav) it dips towards the NW. At GR B 958128 there are poorly exposed examples of horizontal banding. This all implies the presence of an antiformal closure of banding. The axial trace of this antiform is oriented ~065° and the trace appears to correlate with the extension of the Binaniller 3B raft-zone underneath this area, i.e. the banding and the antiformal nature of it may have been produced by syn-plutonic deformation around the raft-zone (see figure 4:48). It would also be logical if the adjacent synform to the NW was produced by similar syn-magmatic shear beneath the mass of early GM 2 tonalite seen in the summit region of Moylenanav. This mass may have controlled the orientation of later granitic sheets as is now seen on the SW slopes of this mountain. Later solid-state deformation with well developed sinistral S-C' fabrics is localised in the antiformal axis of the banding which it also overprints. Away from this zone the C' bands become shorter and more widely spaced until eventually they pass into weakly deformed granitoids typical of this area. The width of this shear zone is 100-125 metres. The location of this zone of high strain may be due to later reactivation of the pelite raft-zone after the surrounding granitoids had crystallised. During the cooling of the pluton the "granite" would progressively become more competent than the pelite. Therefore one might expect high strains to concentrate in the rheologically weaker pelite with such localisation also affecting the immediately adjacent granitoids.

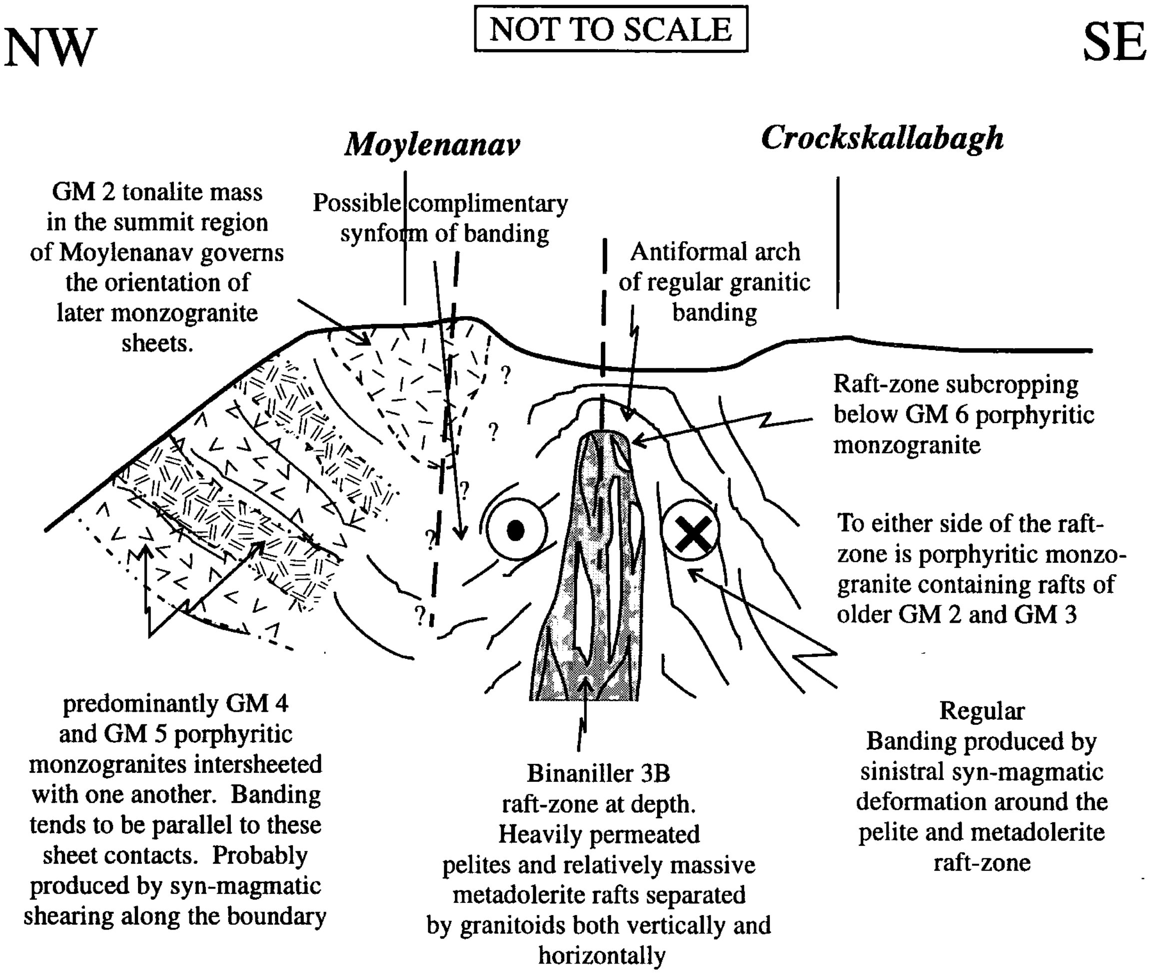


Figure 4:48:- schematic cross-section through Moylenanav-Crockskallabagh (see figure 4:33 for section line), illustrating the orientation of granitic banding in relation to the Binaniller 3B raft-zone and the GM 2 tonalite mass in the summit region of Moylenanav.

4:5:3 Summary of the Granitoids in the Glendowan Mountains area

In the Glendowan Mountains the oldest granitoid phase is the fine-grained, biotite-rich granodiorite, GM 1, which tends to form large pods around which the later granitoids form a crude macrobanding. The pods are preferentially aligned parallel to the foliation and typically have dimensions of 10-20 metres wide by 30-50 metres long and are well exposed on the elongated summit of Croaghacullin. These pods were not observed within the Moylenanav area, although 350 metres to the SE of the Binaniller 3B raft-zone the pods become more common again.

The next phase of intrusion is the equigranular tonalites, GM 2, which are generally coarser than the GM 1 phase. In the Croaghacullin area the GM 2 phase is present but not that common, whilst in the Map II area it is far more abundant forming relatively angular rafts within the younger GM 6 porphyritic host. To the NW of the Binaniller 3B raft-zone GM 2 is less common probably because it has been extensively intruded by GM 4 and GM 5. There is one large remnant of GM 2, which forms the summit region of Moylenanav, and this appears to have governed the orientation of the sheeting of the later granitoids as well as syn-plutonic deformation along the sheet contacts leading to the production of granitic banding.

The more common occurrence of GM 2 to the NW of the GM 1 pods is a relationship analogous to that observed in the Sruhanavarnis area where the SRU 1 granodiorite was seen to be in greater abundance to the NW of the Glenveagh 3A raft-zone than to the SE. In the Croaghacullin area when all the granitoids later than GM 1 are "removed" it is apparent that GM 1 may have been in contact here with the original countryrocks. GM 2 is more abundant to the NW and SE of GM 1 and therefore appears to have preferentially intruded in these areas. As noted earlier, the GM 1 pods were encountered 350 metres to the SE of the Binaniller 3B raft-zone. The occurrence of no GM 1 pods in this intervening ground may be due to GM 2 tonalite preferentially intruding along the interface between the raft-zone and the GM 1 mass, as this would have been a zone of high viscosity contrast. This may suggest that the Binaniller 3B raft-zone was the NW boundary to the GM 1 granodiorite in this area before the intrusion of later granitoids.

The GM 3 facies is a medium-grained aphyric monzogranite which is younger than GM 2. It has essentially similar outcrop relationships to GM 2, i.e. it has intruded earlier granitoids to some degree but occurs as rafts within the later porphyritic GM 4, GM 5 and GM 6 monzogranites.

The porphyritic granites

The exact age relationship between the GM 4, GM 5 and GM 6 monzogranites is uncertain. What is clear is that they all post-date the earlier GM 1, GM 2 and GM 3 granitoids as these are found as inclusions within them. The relatively sharp planar contacts of the earlier aphyric granitoids within the later porphyritic monzogranites imply that there may have been a period of quiescence in this area which allowed the GM 1-3 facies to become quite competent (due to cooling). This feature is in accordance with similar relationships seen between the earlier tonalites and granodiorites in the Sruhanavarnis area and on Crobane Hill.

In Map I, the finer grained GM 4 is seen in contact with the coarser, pinker GM 5 although the exact age relationships are equivocal. On the summit of

Croaghacullin GM 5 was essentially absent whilst GM 4 forms the main host to the earlier GM 1 and GM 3 granitoids. This may imply that the granitoids of this summit area may be a massive composite raft which lies within an intruding GM 5 monzogranite.

On Moylenanav the GM 4 and GM 5 monzogranites are seen together forming low angle sheets inclined to the SE. Along the boundaries of these sheets there is a type of regular type banding, which is morphologically similar to that described by Berger (1967; 1971) in the Doocharry area. This banding, which is well developed on the SW flanks of Moylenanav, appears to be of deformational origin with the bands dominantly composed of GM 4 and GM 5. When traversing from GM 4 into GM 5 the appearance of the banding is as follows:- the homogeneous GM 4 starts to become banded with small, thin bands of the GM 5. Moving towards the GM 5 the GM 5 bands become thicker and the GM 4 bands become thinner. This continues until eventually the GM 4 bands thin and eventually die out into now homogeneous GM 5. This relationship suggests that both these monzogranites may be contemporaneous and were capable of viscous flow at that time, but the viscosity contrast was sufficient to prevent any mixing apart from streaking-out of both facies along the immediate interface due to sub-magmatic deformation.

The relationship of the GM 5 to the coarser white porphyritic GM 6 monzogranite is less certain as contacts between the two were not observed. Overall this later facies is commonly seen in the vicinity of the Binaniller 3B raft-zone, outcropping to either side of it.

The final episode of intrusion consists of a series of microgranodiorite dykes and to a lesser extent pegmatites. The microgranodioritic dykes in appearance and petrography are very similar to the GM 1 granodiorite. Most commonly these small, sometimes sinuous, dykes occur within the granitoids to the NW of the Binaniller 3B raft-zone. The dykes are often discordant to the foliation, (although themselves have been deformed by S_6), and truncate banding, and are seen within all of the porphyritic facies within the Moylenanav area.

The granitoids present in the Glendowan Mountains were looked for in other areas long strike. The ground to the SW of Moylenanav and Crockskallabagh is quite poorly exposed with thick peat deposits forming the Lough Barra Bog. To the NE of the Glendowan Mountains, on the NE side of the Bullaba Valley are the prominent peaks of Farscollop, Mount Kinnaveagh and Leahanmore, the latter of which is situated to the SE of the other two peaks. The granitoids on Leahanmore are essentially homogeneous pink porphyritic monzogranites and will be discussed to a much greater extent in Chapter 5. In contrast, the granitoids on Farscollop and Kinnaveagh are much more heterogeneous and bear strong similarities to those seen in the Croaghacullin area. On these two summits pods of granodiorite resembling GM 1 are quite common, along with raft-like masses of medium-grained tonalite similar to GM 2. The host granitoid to these earlier phases is weakly porphyritic monzogranite which is strongly reminiscent to that of GM 4. The coarser porphyritic varieties (GM 5 and GM 6) were not seen in this area. Quite often around the fine-grained pods regular type banding is well developed, e.g. on Meensnee Hill (southern part of Farscallop) where such banding is very well preserved. In Map D of this thesis the Glendowan Mountains have been divided up into two granitoid packages. The granitoids of the Croaghacullin area and Kinneveagh-Meensnee Hill have been grouped together in the "Glendowan Mountains package". This package is typified by pods of the earlier granodiorite surrounded by tonalite and a dominantly medium-grained monzogranites. The transition between the heterogeneous granitoids of Kinneveagh and Farscollop with the more homogeneous monzogranites of Leahanmore is situated in the low lying valley between Kinnaveagh and Leahanmore.

The northern boundary to the "Glendowan Mountains package" has been conjecturally mapped as the point where fine-grained granodiorites disappear with the dominant granitoid types being medium-grained tonalite and a suite of often very coarse porphyritic monzogranites. The coarser nature to these monzogranites may be due to slow cooling rates within the central regions of the pluton with the ambient temperature having risen due to the intrusion of the earlier more biotite-rich and finer grained porphyritic monzogranites. These granitoids have been grouped together to form the "Doocharry-Moylenanav package". A common feature of this package is regular granitic banding which is seen in the far SW (along SE side of Gweebarra River), in the Doocharry area and also in the Moylenanav area. It is possible that this package has subjected to considerable syn-magmatic deformation.

4:6 Discussion

This section will address the heterogeneous zone as a whole within the central regions of the Main Donegal Granite and will attempt to give an insight into the evolution of this region of the pluton. Table 4:1 summarises the chronological order of the granitoids in the three areas studied (Crobane Hill, the Sruhanavarnis Valley and the Glendowan Mountains).

The overall sequence of emplacement is similar in all areas, although in some areas extra phases are present. The early phases, which tend to be equigranular tonalite and granodiorites, all occur as raft-like masses within the later and generally more abundant porphyritic monzogranites. The contacts of these early granitoids with their porphyritic host are quite sharp, at least on the outcrop scale, and implies that these early granitoids were generally quite competent during the intrusion of the later porhyritic monzogranites.

Crobane Hill	***Sruhanavarnis Valley***	***Glendowan Mountains***
? CBH 4*: fine-grained, equigranular tonalite	SRU 1: fine to medium-grained, equigranular granodiorite	GM 1: fine-grained equigranular granodiorite
CBH 1: medium-grained equigranular granodiorite/ tonalite	SRU 2: medium-grained equigranular biotite-tonalite	GM 2:medium-grained, equigranular biotite-tonalite
CBH 2; medium-coarse porphyritic monzogranite	SRU 3: porphyritic monzogranites. Possibly medium-grained PSBG, PSBGp ↓ Medium to coarse-grained PPBG ↓ Coarse-grained VCPG and CWBG	GM 4: medium-grained, subtly porphyritic monzogranite ↓ Coarser porphyritic monzogranites Pink: GM 5 White: GM 6
CBH 3: medium-grained equigranular biotite-poor monzogranite		
CBH 4 dismemebered microtonalite dyke	Early tonalite syn-plutonic dykes (Early transverse dyke of Pitcher & Read (1959)	Fine-grained tonalite-granodiorite dykes resembling GM 1 in appearance
CBH 4: regular microgranite dykes	Later, regular microgranite dykes	
Late pegmatites	Late pegmatites	Late pegmatites

Figure 4:49:- Table summarising the chronology of granitoid emplacement in the three areas of detailed mapping: Crobane Hill, Sruhanavarnis and the Glendowan Mountains.

Contact relationships between these earlier tonalites and granodiorites are less commonly observed due to later monzogranites appearing to preferentially intrude along pre-existing contacts. The distribution of the tonalites and granodiorites in the central regions of the pluton leads the author to suggest that this may be the oldest part of the pluton (see chapter 8). It is probable that the Main Donegal Pluton was not constructed in one continuous event but episodically. The porphyritic monzogranites, which have disrupted this possible earlier more tonalitic and granodioritic pluton, appear to have been emplaced quite rapidly due to the lack of clear contacts between the different porphyritic types.

A question arises to the extent of this heterogeneous zone within the entire pluton. To the NW of the Derryveagh 2 raft-zone, which is the north-westerly boundary of the heterogeneous granitoids, there are very homogeneous porphyritic monzogranites. At the NE end of the pluton, on the south-eastern margin quite homogeneous granitoids are again found and are believed to be younger than the heterogeneous zones to the immediate NW. Therefore at this NE end of the pluton the heterogeneous granitoids are sandwiched between essentially homogeneous and possibly later monzogranites. These homogeneous zones form the basis of Chapter 5 and will not be addressed further here. To the SW of this area the heterogeneous zone gets wider as the major raft-trains diverge. Much further to the SW beyond Crobane Hill, Sruhanavarnis Valley and the Glendowan Mountains the southerly margin of the pluton is composed of dominantly heterogeneous granitoid with the homogeneous zones observed at the NE end of the pluton being absent.

Chapter 5

The homogeneous granitoid zones in the more marginal regions of the Main Donegal Granite

5:1: Introduction

Within the Main Donegal Granite there are zones of more homogeneous monzogranite which are characterised by the general absence of country-rock meta-sediments and other granitoid facies. Within each of these zones contacts are very rare and variations tend to be highly transitional in nature. The boundaries to these zones are generally abrupt and marked by the presence of country rock raft-zones and/or septa which strongly suggests that the country-rock architecture has influenced the spatial distribution of intruding granitoids. Granitoids of this type were studied in detail in an attempt to define sheet boundaries based on mapping in the Barnes Gap-Crockmore area (NE end of pluton). Along strike to the SW (Losset and Leahanmore-Croaghacormick) the granitoids were studied in an attempt to identify any correlation with the Barnes Gap area. Further to the SW the homogeneous granitoid zones tend to pass into more heterogeneous zones and granitoids typical of the country rock boundary zones; "*the marginal facies*" (i.e. pegmatite and banded granodiorites with variable biotite contents).

In the NW of the pluton homogeneous porphyritic monzogranites were observed to the north of the Derryveagh 2 raft-zone although on the immediate NW margin of the pluton the marginal facies are again seen. This homogeneous monzogranite of the NW was studied in the Lackagh Bridge-Cocks Heath Hill area, Dunlewy-Poisoned Glen area, at Kingarrow and in the Meenderryherk-Brockagh area. This overall relationship is illustrated in Figure 4:1 of the previous chapter and Map D of this thesis.

5:2: The homogeneous granitoids of the SW margin (NE end of pluton)

Along the SE margin two zones of homogeneous granitoid are encountered. The south-eastern zone is an aphyric granodiorite-monzogranite which is internally very homogeneous. To the NW of this zone there are homogeneous pink, weakly porphyritic monzogranites which show internal grain size variations, although no major modal differences have been encountered. The boundary between these two zones is marked by a discontinuous zone of rafts which make up the Crockmore 4 raft-zone. In the Barnes Gap-Crockmore area these granitoids are well exposed as are the meta-sedimentary partitions which separate the two zones; thus making this area suitable for detailed study.

5:2:1 Barnes Gap-Crockmore

Figure 5:1:- The well exposed granitoids of the Barnes Gap looking NE from the NW flanks of Stragraddy Mountain. The prominent break of slope on the left is the Binaniller Hill. The escarpment in the distance on the left is that of Crockmore and Crocknacrady.

In this area the granitoids were studied to the NE of the N56 road in a traverse from the old railway bridge at GR C 081264 to the SE margin of the pluton at GR C 093242. The mapping area included the high ground around the Crockmore summit itself extending north-westwards to the small hill of Derriscligh (GR C 091269). Within the Barnes Gap there are numerous raft-zones which in comparison to other areas within the pluton, are very closely spaced. Further to the SW these become progressively more separated. The most prominent raft-zones in this area are the

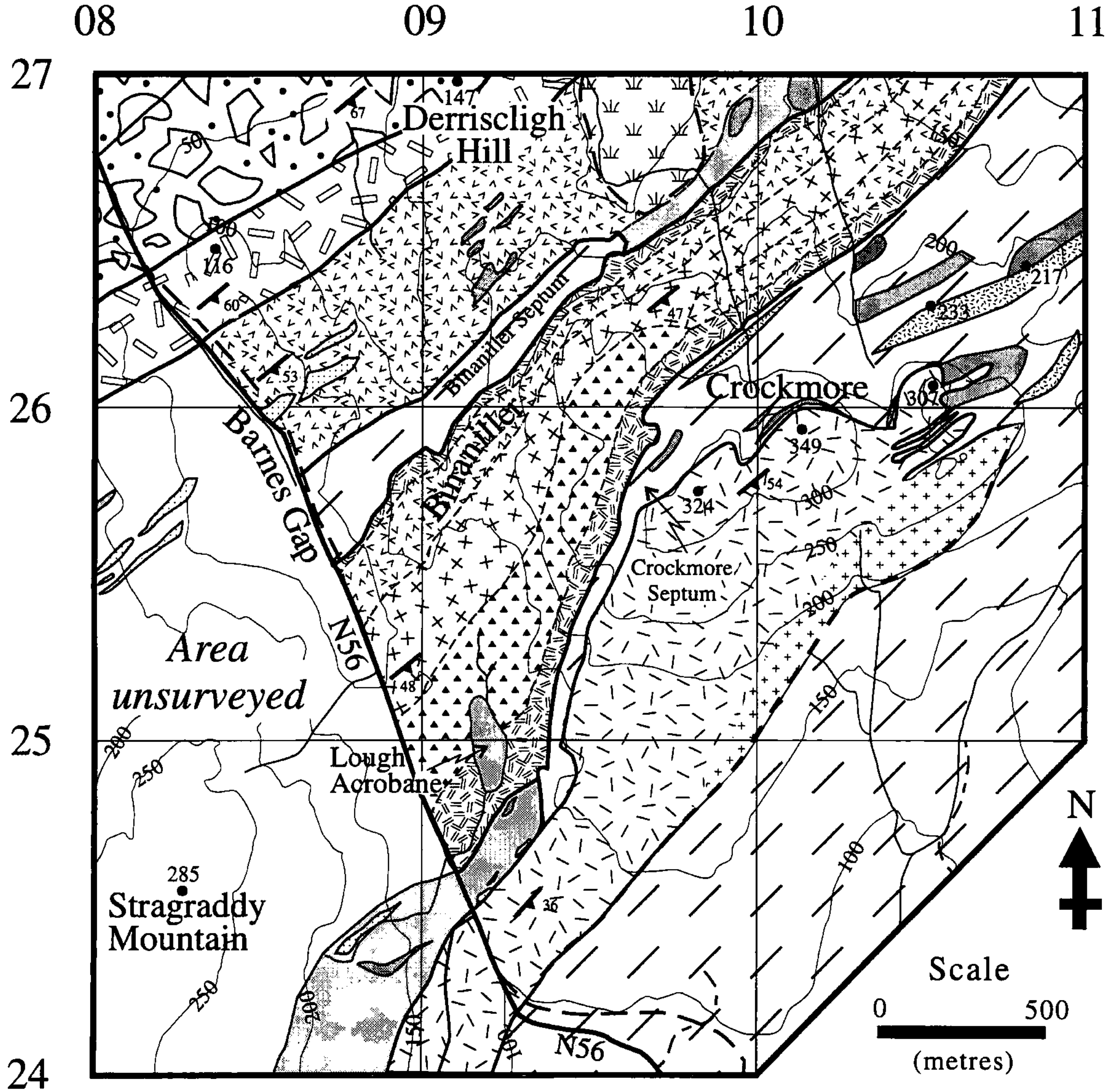


Figure 5:2:- The Granitoids of the Crockmore - Barnes Gap area, NE Main Donegal Granite

Legend

- White aphyric biotite-muscovite monzo-granite (Crockmore Granite)
- Muscovite-garnet aplogranite
- B.a .1 White aphyric biotite granitoid
- B.a. 2 Coarse-grained
- B.a. 3 Medium-grained
- B.a. 4 Fine to medium-grained

(B.a. 2–4: Pink porphyritic monzogranites)

- Heterogeneous granitoids rich in quartzite xenoliths
- Heterogeneous granitoids rich in Thorr xenoliths
- Metadolerite
- Microgranite sheets
- Disrupted septa (rafts "floating" in granite)
- Sheeted country-rock
- Pelite rafts
- Peat Bog or poor exposure

Crockmore Septum 4 and the Binaniller 3B raft-zone. These two raft-zones dip moderately to gently to the SE with the granitoids above both of the raft-zones tending to be relatively homogeneous. The monzogranite which lies above the Binaniller 3B raft-zone and below the Crockmore Septum will be called the Binaniller apophysis, whilst the equigranular monzogranite-granodiorite above the Crockmore septum will be termed the Crockmore apophysis. The granitoids to the NW of the Binaniller 3B raft-zone are compositionally more heterogeneous with contacts between the different granites facies generally quite common. Within these granitoids are scattered xenoliths of quartzite (and to a lesser degree pelite) whilst at the NW end of the Barnes Gap xenoliths of Thorr Granodiorite become numerous. It is these two xenolith types which form the prominent Derryveagh 2 raft-zone which has already been described to some degree in earlier sections (4:2 and 4:3) of this thesis.

The overall distribution of raft-zones and different granitoid facies is illustrated in figure 4:1, figure 5:2 and Map D of this thesis.

5:2:1:1 The Crockmore Septum and overlying apophysis

The Crockmore Septum is dominantly composed of pelite, meta-dolerite and calc-silicate which are believed to belong to the Upper Falcarragh Pelites (Pitcher & Berger 1972). These meta-sediments have been intruded by a "plexus" of granite sheets which show wide variation in size and composition (aplogranites to biotite granites). The outcrop width of the septum in the Crockmore summit area is 250 metres whilst to the SW this value decreases until eventually it breaks up in to smaller rafts, 100 metres to the east of Lough Acrobane. Mapping reveals that the septum is reduced in thickness primarily by the cross-cutting nature of the overlying monzogranite (the Crockmore Apophysis), and to a lesser extent, by the cross-cutting nature of the lower Binaniller apophysis. The septum is gently inclined to the SE and on average has dip angle between 30-40° and forms the base of the escarpment on the NW side of Crockmore (see figure 5:3).

The dominant granitoid immediately above the septum is a very homogeneous light-grey, biotite-rich aphyric monzogranite-granodiorite and forms a prominent escarpment above the septum clearly visible from the Barnes Gap (see figure 5:3). The approximate thickness of the sheet in the Barnes Gap railway cutting is ~ 215 metres, (assuming a dip of 35°). The thickness of this sheet in the Crockmore area is less certain due to the poor exposure on the dip side of the summit. Above this homogeneous sheet of monzogranite are a series of smaller sheets which are clearly separated by intercalations of pelite. Figure 5:4 shows these sheets from the summit of Crockmore. On the left in the middle ground is the main Crockmore apophysis. To the SE (right) there are two small sheets of equigranular biotite granitoid (grain

Figure 5:3:- The Crockmore Septum viewed from the SE (the N56 road). Pelite can be seen at the base of the escarpment with the Crockmore apophysis lying above. On the far left are granitoids belonging to the Binaniller apophysis.

Figure 5:4:- Smaller granitoids sheets above the main Crockmore Sheet. Looking NE from the summit of Crockmore (see text).

size and texture differ from that of the main apophysis). Finally further to the right is the uppermost sheet which is composed of muscovite-garnet aplogranite (contains low amounts of biotite).

The SE contact of the Main Donegal Granite in the Barnes Gap area is only clearly visible in the old rail-cutting. Figure 5:5 shows a section across this contact over a distance of 100 metres. The first 50 metres at the NW end of the section consist of homogeneous pale-coloured biotite-monzogranite of the Crockmore apophysis although it is generally coarser in this area compared to lower parts of the sheet. The SE contact of the Crockmore sheet is very sharp and almost vertical and it clearly cuts the foliation ($S_{2/3}$ of Berger 1967) within the host pelite. As commented in Pitcher & Berger (1972) the main "granite" also truncates a small folded quartzo-feldspathic vein although the exact age relationship of this fold was uncertain. The remainder of the section shows the sheeted nature of the country rock-pluton contact with pelite and metadolerite intruded by a suite of low angle sheets (commonly 25-35°) which are composed of microgranite and pegmatite. The microgranites differ from one another in grain size and colour, the latter reflecting the proportions of biotite and muscovite. Along this section there were no clear examples of minor intrusives cross-cutting one another nor the Crockmore apophysis cutting the marginal sheets. The garnet-muscovite aplogranite sheet seen above the Crockmore apophysis in the Crockmore area is not in evidence in the Barnes Gap implying that it lenses out in the intervening ground.

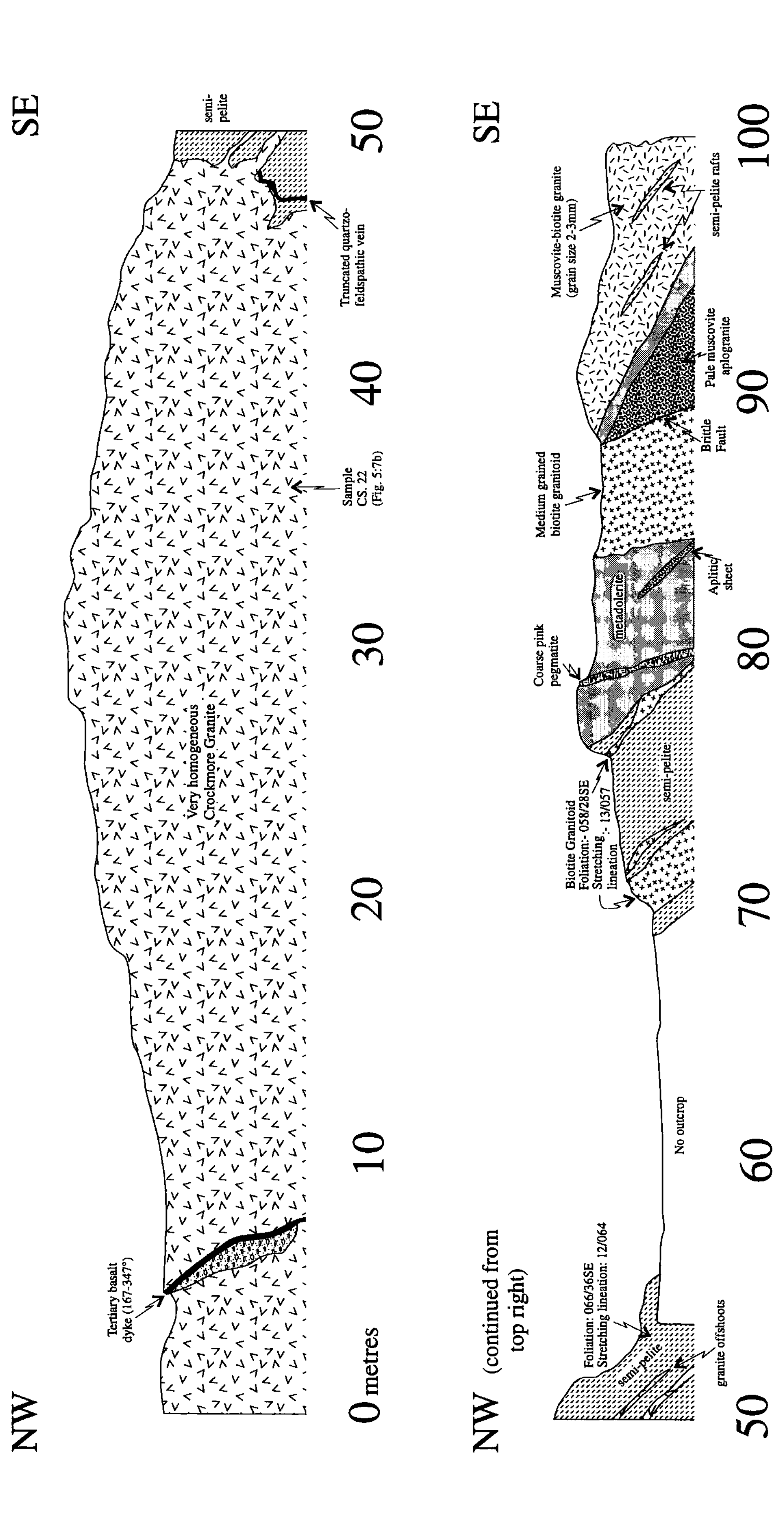
Deformation within the Crockmore apophysis

On the whole the majority of the Crockmore apophysis is only weakly deformed with a weak to moderately strong S_6 foliation dipping gently to the SE (see figure 5:6 stereonets). Near the contact with the Crockmore pelitic septum and at the main contact of the pluton the deformation greatly intensifies with the development of strong sinistral S-C' fabrics (figure 5:7a). In the cross-section through the contact the strain only notably intensifies within the immediate pelites at the margin of the pluton. The orientation of the S_6 and S_8 (C') foliations implies inclined sinistral shearing (X plunges shallowly to the NE) acting on the pluton in this area.

Thin-Section

A specimen collected 15 metres from the contact (see figure 5:5 for location) shows relatively low strain (figure 5:7b) in comparison to the marginal sheets and the immediately adjacent envelope rocks. Red-brown biotite shows only a moderate alignment in thin-section with minor secondary kinking. Plagioclase and microcline are essentially undeformed apart from strain induced myrmekite in the latter mineral.

Figure 5:5:- Cross-Section through the rail-cutting at the SE end of Barnes Gap showing the contact and sheeted margin of the Main Donegal Granite



Quartz is the only mineral which displays evidence of ductile deformation and shows varying degrees of undulose extinction. The presence of subgrains and deformation lamellae implies there has been minor recovery by dislocation "climb" with smaller undeformed grains recovered by dynamic recrystallisation (both sub-grain rotation recrystallisation (SR) and grain boundary migration recrystallisation (GBM) (after Passchier & Trouw 1996)).

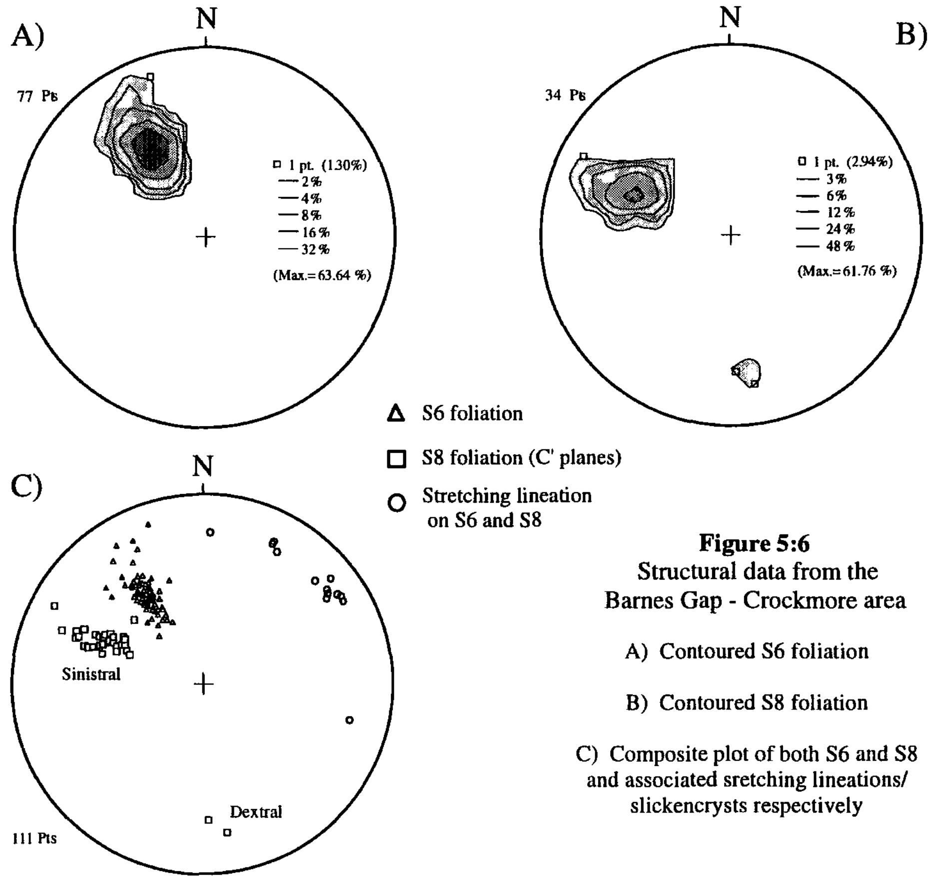


Figure 5:6
Structural data from the Barnes Gap - Crockmore area

A) Contoured S6 foliation

B) Contoured S8 foliation

C) Composite plot of both S6 and S8 and associated sretching lineations/ slickencrysts respectively

The sample shown in figure 5:7c is Crockmore monzogranite taken from above the septum (10-20 metres). Strain is much higher within this area of the apophysis (protomylonitic granite) with the development of S-C' fabrics. The grain size of this rock has been considerably reduced by deformation compared to the same monzogranite in the rail-section. The foliation is marked by the preferred alignment of biotite and muscovite which lie along both the S and C' foliations. Grain size reduction in quartz intensifies into the C' planes with dynamic recrystallisation processes producing relatively unstrained grains. The size of the recrystallised grains is a function of differential stress (Passchier & Trouw 1996) i.e. the higher the differential stress the smaller the grain size. Feldspars tend to behave as porphyroclasts with this mineral phase tending to be more resistant and display

Figure 5:7a:- Crockmore Granite immediately above the septum showing intense sinistral deformation.

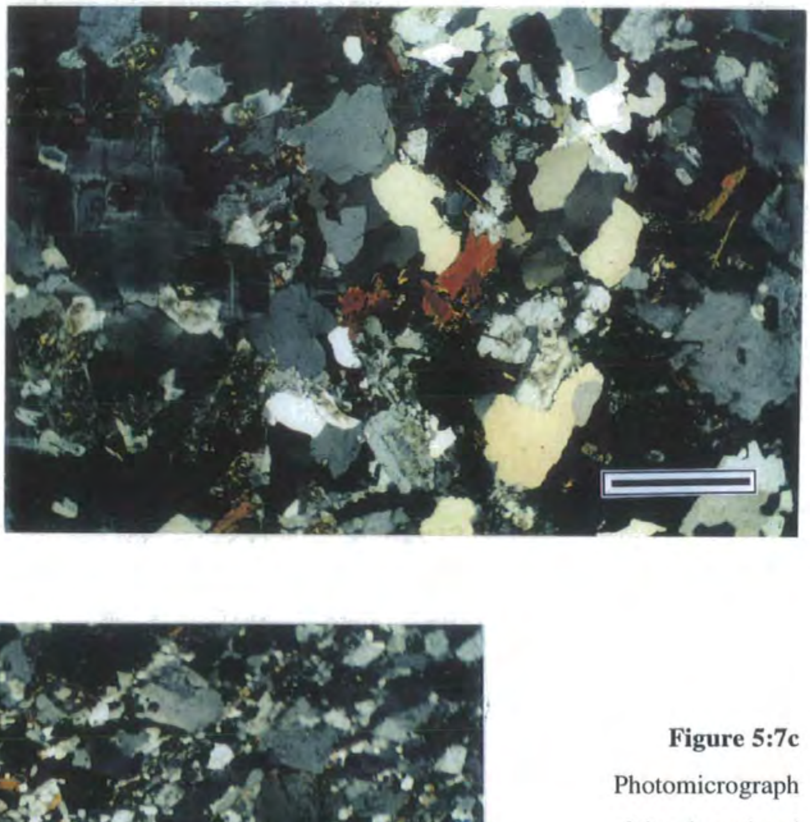


Figure 5:7b
Weakly strained Crockmore Granite from the rail-cutting near the margin of the pluton (scale bar =1mm).

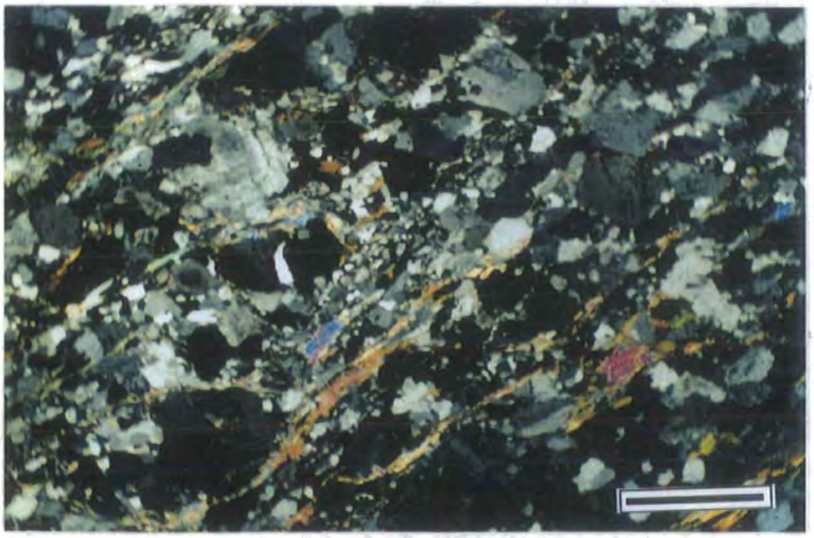


Figure 5:7c
Photomicrograph of the above hand specimen (fig. 5:7a) (scalebar = 1mm).

dominantly brittle deformation by cracking. The spacing of the C' planes is generally less than 1 mm (this spacing is governed by the size of the feldspars), whilst the lengths of these planes is greater than that of the section (>3 cm). From both hand specimens and thin-sections the sense of shear is consistently sinistral. The orientation of these cleavages indicates inclined ductile sinistral shearing, with stretching lineations plunging in a north-easterly direction at 15-25°.

5:2:2 The Binaniller 3B raft-zone and overlying apophysis

The Binaniller 3B raft-zone is dominantly composed of pelite, meta-dolerite and limestone and forms the base to the NW slope of the hill called Binaniller. In figure 5:1 the photograph shows a view of the Binaniller apophysis from the NW flanks of Stragraddy on the SE side of the N56 road. The well exposed granitic material in the central portion of the photograph belongs to the Binaniller apophysis and forms the majority of the exposure in the Barnes Gap. The topographically highest escarpment on the right is the Crockmore apophysis with the Crockmore septum forming the depression immediately in front of it with the top of the Binaniller apophysis immediately below.

In the old rail-cutting at GR C 086257 limestone rafts of the Binaniller 3B raft-zone are well exposed, inclined (40-45°) to the SE. Traced in a NE direction from this rail-cutting are abundant rafts of pelite which become continuous enough to be classed as a septum. Between GR C 089258 to GR C 094262 this pelitic septum can be traced along the NW base of Binaniller although it is often covered with granitic boulders. The granitoid-pelite contact at the top of the Binaniller septum is oriented ~040°/42°SE. The pelite shows minimal alteration (i.e pegmatisation) due to the surrounding granitoids. Further to the NE pelite and metadolerite rafts (clearly surrounded by granitic material, i.e. no longer a septum) can be traced along the steep south-eastern boundary of the large bog. A large metadolerite mass at GR C 102269 marks the north-eastern limit of the Binaniller 3B zone of meta-sediments and beyond this, no more rafts of pelite and metadolerite are encountered. The pelites and metadolerites which form this septum or raft-zone have been intruded by a suite of compositionally different granitoids (although the most common variety is a white equigranular, biotite-rich granitoid) in a similar fashion to the Crockmore Septum. These sheets are usually concordant to the schistosity in the countryrock (see figure 5:8). The dip of the upper contact (SE side in fig. 5:2) of the Binaniller apophysis appears to differ from that of the base. It appears to be almost vertical as indicated from mapping the stream sections to the north of Crockmore. A similar feature can be seen in figure 5:5 where the upper contact of the Crockmore apophysis is almost vertical and locally discordant to the schistosity in the host pelites.

Figure 5:8:- photograph of small biotite-rich, equigranular granitoid intruding into the Binaniller septum (GR C 094264).

The monzogranites which form the Binaniller apophysis are generally very homogeneous with no internal contacts, or any xenoliths of meta-sedimentary or earlier granitic rafts within it. Variation was seen along the upper and lower contacts where a white, biotite granitoid is locally visible. The majority of the apophysis is a pinkish-white porphyritic monzogranite which shows internal variations in grain size which were mapped (see figure 5:2). One must emphasise that the author expended much time in trying to find contacts between these varying monzogranites but on outcrop scale the contacts tend to be highly transitional. The variations within the Binaniller apophysis are as follows:-

i) B.a. 1:- immediately to the NW (or below) the Crockmore Septum is an aphyric medium to coarse-grained (3-4 mm) biotite granitoid with white feldspars (see appendix A). It is coarser than the Crockmore Granite and overall is less deformed than it. This granitoid can be traced the entire strike length of the mapping area with its outcrop width rarely exceeding 200 metres. Along the base of this apophysis, immediately above the Binaniller (3B) septum-raft-zone, a similar biotite granitoid with white feldspars is also visible although the grain size and biotite contents are more variable. These variants can be observed in the two streams which flow northwards from either side of Crockmore where the different facies are well exposed in gorges. Granitic material resembling B.a. 1 was also seen in some locations within

the pink porphyritic monzogranites, in the central areas of the apophysis, where it tended to form large homogeneous masses; although no contacts were observed.
ii) B.a. 2:- this is the coarsest monzogranite within the Binaniller apophysis and owes this feature to its large pink microcline megacrysts (~5 mm) set in a finer 2-3 mm groundmass. This monzogranite outcrops to the NW of the B.a. 1 and forms the high ground 0.5 km to the west of Crockmore. It is also visible from the roadside of the N56 on the small hill NW of Lough Acrobane (GR C 090251). To the NE of GR C 098262 the porphyritic monzogranites become finer grained. No sharp contact is visible and the boundary is transitional over a distance of approximately 100 metres.
iii) B.a.3:- this is finer grained than B.a. 2 but petrographically similar having a grain size 1-2 mm with 3-4 mm microcline megacrysts. This phase appears to contain more biotite than B.a. 2 but this may be an apparent feature as a result of smaller K-feldspar megacrysts. In the Barnes Gap, nearer to the road (N56) the porphyritic monzogranites have a strong red colour probably caused by subsequent alteration (i.e. faulting or the intrusion of later Tertiary basalt dykes. In the north-eastern area of figure 5:2 this medium-grained monzogranite forms the central part of the Binaniller apophysis.
iv) B.a. 4:- the finest grained member of the apophysis (1-2 mm groundmass, 2-3 mm microcline megacrysts) tends to occur lower down in the apophysis although in the NE it lies to either side of the B.a. 3 granite. In this latter situation the grain size could be the result of chilling as the thickness of the apophysis is the least in this area.

On the whole the Binaniller apophysis is more intensely deformed than the Crockmore apophysis. The strain within the Binaniller apophysis increases towards both of the septa. In parallel with a decreasing grain size and with a reddening of the granites towards the contacts of the apophysis. These relationships introduce the possibility that grain size reduction due to chilling has been exploited during deformation with higher strains localised here together with fluid ingress.

Deformation

Outcrop:- as already mentioned, the granitoids within the Binaniller apophysis are heavily deformed with the development of a pervasive S_6 foliation. The inclination of S_6 has increased, in comparison to the S_6 in the Crockmore apophysis, and has a typical dip angle of 40-50° to the SE. Towards the margins of the apophysis this is overprinted by strong S-C' fabrics which show consistent inclined sinistral shear. Present to a lesser extent are antithetic C' fabrics oriented approximately E-W. Despite strain being lower overall in the centre of the apophysis, zones of high strain are observed, e.g. GR C 090251. The deformation within the Binaniller septum is very intense with S-C' fabrics similar to that seen in the adjacent apophysis, present in both pelite and metadolerite, accompanied with a well formed stretching lineation (indicated by acicular hornblende in the latter), which in this local area consistently plunges gently towards the NE.

Thin-section

Within all of the thin-sections looked at there is considerable deformation-related grain-size reduction, most notably along the C' planes. The mineral constituents will be addressed separately.

1) Plagioclase:- this has behaved predominantly in a brittle fashion with fractures present, approximately at right angles to the albite twinning, which have been infilled with quartz. There is a possibility that this feature is related to sub-magmatic microfracturing with the cracks infilled with quartz or K-feldspar melt (Bouchez *et al.* (1992) although one must emphasise that these monzogranites have been overprinted by strong solid-state deformation. The presence of plagioclases with "sweeping" extinction does not indicate that this mineral is deforming in a ductile fashion but is due to sub-microscopic brittle fracturing within the crystal (Passchier & Trouw 1996).

2) Microcline:- despite having similar rheological properties (Passchier & Trouw 1996) the microcline within these monzogranites shows signs of grain size reduction. Again cracks which have been infilled with quartz are quite common indicating possible sub-magmatic deformation. Solid-state strains are shown by a thin "rind" around some large megacrysts (porphyroclasts) which is very fine-grained (<100μm) generally much finer grained than the recrystallised quartz. This feature implies that the microcline was starting to deform in a ductile manner with dynamic recrystallisation occurring along the margins (see figure 5:9). Furthermore the presence of weak undulose extinction also suggests that dislocation creep processes may have been operating. Strain-induced myrmekite is very common and also forms very fine-grained aggregates around microcline.

3) Quartz:- this mineral shows a variety of phenomena within these granites. The quartz grains which are furthest away from the S and C' planes show strong undulose extinction and inhomogeneous flattening. Quartz which lies along these two planes is fine-grained and has recovered by SR and GBM processes with these grains only showing weak undulose extinction. In close proximity to feldspar grains quartz has been strongly flattened with aspect ratios as high as 15:1. Finally within these monzogranites there are zones of weakly deformed crystals of quartz which appear to have completely strain "healed". The presence of planar boundaries and 120° triple junctions implies strain healing has occurred, possibly due to deformation ceasing with subsequent static recrystallisation.

4) Biotite:- this is responsible for producing the macroscopically obvious foliations within these monzogranites lying along both the S and C' planes. This mineral phase tends to be more deformed around feldspar porphyroclasts. Muscovite tends to have similar geometric relationships as biotite.

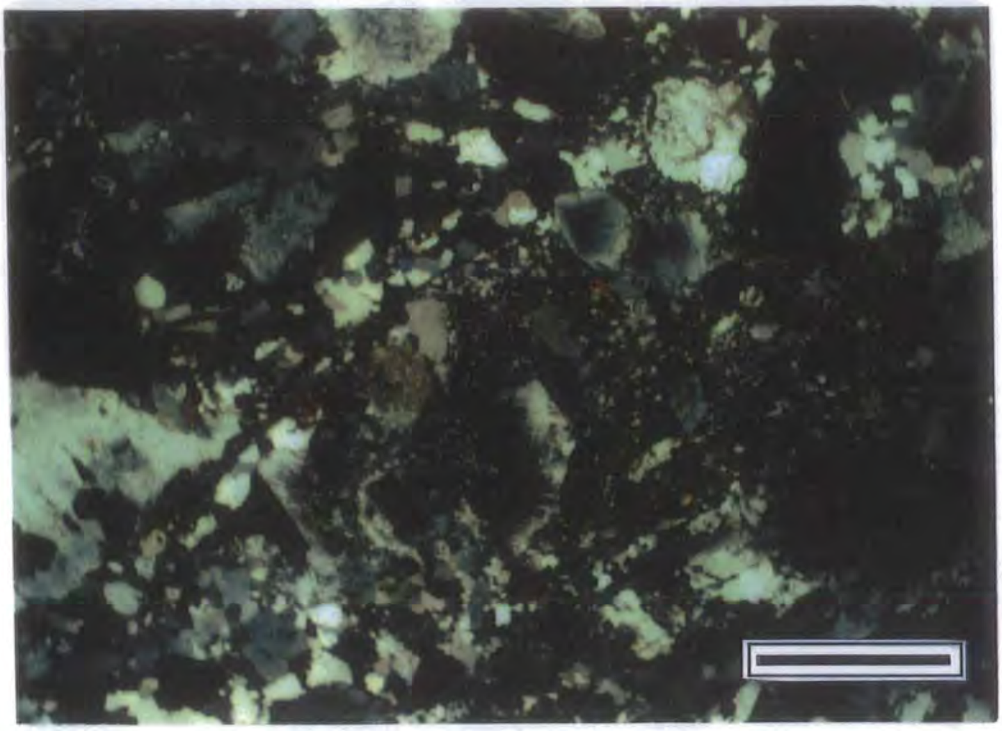


Figure 5:9:- High-temperature solid-state deformation within the B.a. 3 monzogranite. Note the very fine-grained mantles of recrystallised K-feldspar around microcline cores. Quartz is also recrystallised with relatively small grains having a strain-free appearance. Larger palaeoblasts of quartz tend to have strong undulose extinction.

The above evidence from thin-section implies these granitoids have undergone sub-magmatic state deformation indicated by weak alignment of plagioclase, isolated occurrences of tiling and feldspar cracks. This sub-magmatic deformation has been followed by high temperature solid-state deformation, generally below the brittle-ductile transition in feldspars (i.e. 400-500°C). This later deformation is very intense and has obliterated much of the earlier magmatic fabrics.

5:2:3 Granitoids to the NW of the Binaniller septum-raft zone

To the NW of the Binaniller septum the percentage of exposure drastically decreases which makes contact relationships between different granitoids more difficult to decipher. Parallel, and to the NW, of this septum are a series of large pelite (GR C 087261) and metadolerite rafts (GR C 094263). These rafts have dimensions of up to 100 $\times$ 40 metres and are clearly "floating" within granitic material. Between these rafts and the Binaniller apophysis is up to 120 metres of homogeneous red-pinkish porphyritic monzogranite resembling B.a. 3. At GR C 100269 a similar granitoid sheet up to 80 metres wide was seen separating pelitic and metadoleritic material. For 200 metres the NW of the above mentioned pelite and metadolerite rafts the granitoids are relatively homogeneous (resembling B.a 3) although beyond this the granitoids become very heterogeneous From GR C 083262 to the old rail bridge at the NW entrance to the Barnes Gap (GR C 082263) the granitoids contain pelite with quartzites becoming more common towards the NW. In the rail-cutting near the old rail bridge these xenoliths are surrounded by up to four different granitoid facies. The presence of orange-brown ochreous waters in this section, and the poor exposure in the area to the immediate NE, prevents any age relationships from being ascertained. Pelite xenoliths (GR C 083265) here show the highest degree of permeation by the surrounding granitoids in the Barnes Gap area and are partially assimilated (see figure 5:10b) whilst quartzite and metadolerite appear unaltered. Pitcher & Read (1959) attributed this to the presence of the Thorr Pluton in this area before it was itself intruded by the Main Donegal Granite with the pelite being heated again by this later pluton. Seventy-five metres to the NW of figure 5:11(a) xenoliths of Thorr Granodiorite (TG) become very common with the transition from quartzite to Thorr diorite being very abrupt. A similar abrupt transition is also visible on the summit of Derriscligh Hill, to the NE (GR C 091269). The larger TG xenoliths (1-2 metres in length) are angular, tending to be tabular with their long axes aligned parallel to the foliation within the host granitoid whilst the smaller TG xenoliths are strongly streaked out. The TG and quartzite xenoliths make up the Derryveagh 2 raft-zone which is up to 1 km wide at this end of the pluton, although in this area most of it occurs in the poorly exposed Owencarrow valley.

The deformation within the granitoids between the Binaniller septum and the Derryveagh 2 raft-zone is generally quite low with only a moderate S_6 foliation developed. In the vicinity of the Derryveagh 2 raft-zone though the strain notably increases with xenoliths of TG having developed S-C' fabrics with the granodiorite having an augen appearance due to large plagioclase phenocrysts being "wrapped" around by biotite. The strain within the host MDG facies is also quite high in this area with less prominent S-C' fabrics developed. Along the Newbridge-Glen road 0.5 km to the NW of Derriscligh Hill the deformation is very intense with well developed stretching lineations on the S_6 foliation and slickencrystic mica on the penetrative C' (S_8) planes. On both these surfaces the plunge of the lineation and slickencrysts is to the NE and NNE respectively. In the Newbridge area the foliation has again steepened with it commonly having an inclination to the SE of 55-65°. Figure 5:10(b) implies there may have been an external control on fracturing during the emplacement of some of the Main Donegal Granite facies. The orientation of these granitoid-filled fractures is 020°, an orientation consistent with that of shear bands developed throughout the pluton, which in this case also show sinistral offsets. Another situation where the host granitoid has behaved in a more ductile (i.e. viscous) fashion is seen in figure 5:10(c) where a metadolerite raft has been brittlely deformed by antithetic fracturing. The foliation in the adjacent granitoid host is deflected around the metadolerite with S-C' fabrics showing sinistral offsets.

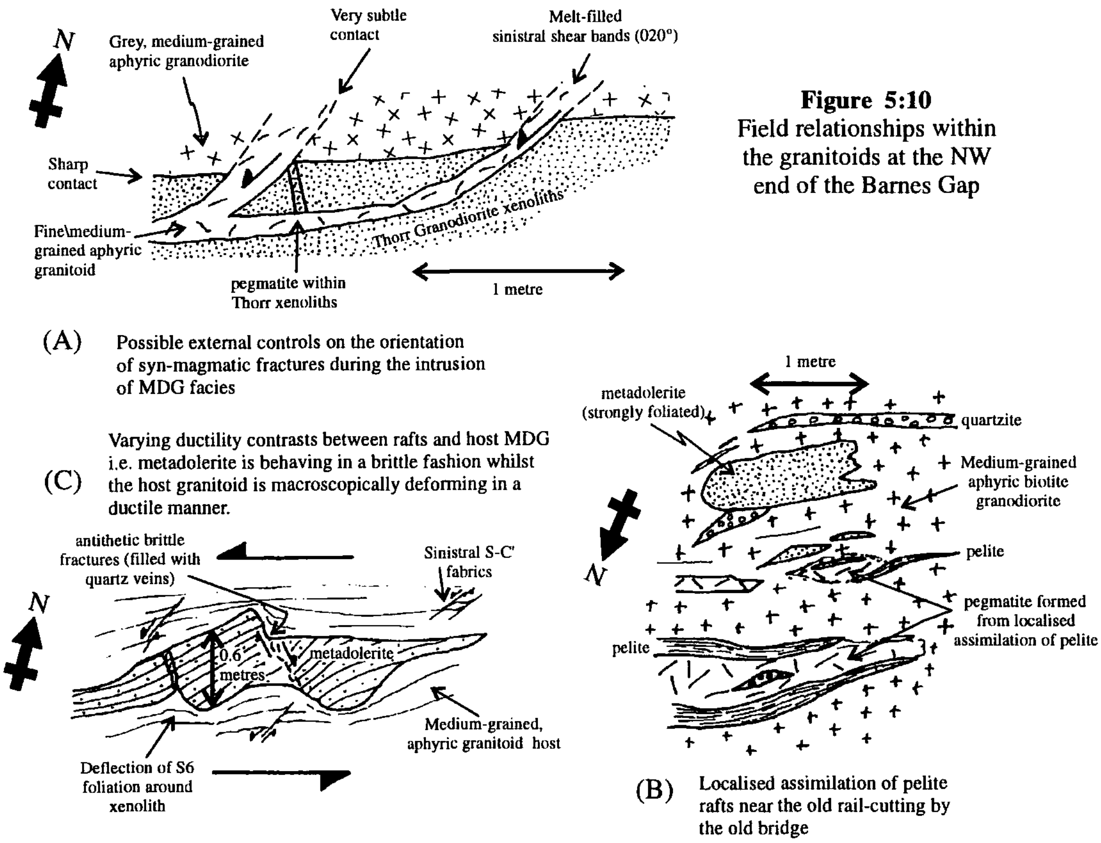


Figure 5:10
Field relationships within the granitoids at the NW end of the Barnes Gap

5:3:4 Summary of the Crockmore-Barnes Gap area

The granitoids in this area are generally quite homogeneous in comparison to those observed in the medial regions of the pluton; and this is especially the case with the monzogranites which lie above, to the SE, of the Crockmore septum and the Binaniller septum. To the NW of these two apophyses the granitoids become more heterogeneous with xenoliths of meta-sediments and older igneous material being common. Despite a lack of unequivocal evidence it is believed that these two apophyses are younger (especially the Crockmore Sheets) than the granitoids to the NW. The reasons are as follows:

1) The granitoid facies which comprise the apophyses between these septa are very homogeneous and contain almost no country-rock xenoliths nor any sharp contacts with any other granitoid facies.

2) If these monzogranites were of early origin then one would expect to see sheets of smaller compositionally variable granitoids which are so common along the contact zones and within the septa and raft-zones, intruding into them.
3) The Crockmore sheet as a whole is generally weakly deformed as Tozer (1955) correctly states. At the contact of the Crockmore sheet (i.e. Main Donegal Granite "proper" contact) this equigranular monzogranite is very weakly deformed in contrast to the intensely deformed schists and marginal sheets.
4) Tozer (1955) also believed the Crockmore sheet to be younger than the main pluton mass to the NW, based on the presence of muscovite porphyroblasts within the Crockmore Septum which are commonly oblique to the foliation and become coarser in grain size towards this monzogranite. Tozer (1955) attributed this to the greisening of the country-rocks by a *"late stage, volatile-rich, differentiate of granitic magma"*. Furthermore the random nature of the muscovite porphyroblasts implied to him an almost static origin, a feature which the present author would dispute since the granitoid in question, immediately above the septum, is intensely deformed in the solid-state.

It is therefore believed that the two relatively homogeneous apophyses were emplaced into country-rock which had already been intruded by a plexus of smaller granitoid sheets. The septa containing the sheets are therefore the remnants of this country-rock material. The reason why there is a lack of country rock or older granitoid inclusions within the homogeneous monzogranites is a feature which needs to be addressed. One possibility is that the buoyancy effect and possible magmatic "head" of these larger masses simply wedged the septa apart with little stoping at the present level of exposure. Another reason may be that the homogeneous monzogranites intruded slightly higher in the crust than the heterogeneous granitoids to the NW and that their higher parts, possibly in the "stoping zone" have been removed by erosion. These matters will be discussed more generally in later parts of this thesis (chapter 8).

The Binaniller and Crockmore septa are intensely deformed and appear to have localised strain. The granitoids to the either side, most notably the Binaniller apophysis, also record high-temperature solid-state deformation. It would appear logical that strain was localised into the septa when the competency contrast between the cooling "granite" and pelite was much lower. The "granite" would eventually become mechanically stronger than the pelite and hence the strain would localise within these septa. This feature would also account for the strain being most intense within the immediate aureole of the pluton. A similar feature, where strain localises into raft-zones was also seen in the Binaniller 3B on Moylenanav-Crockskallabagh in chapter 4 (section 4:4:2).

The Binaniller apophyses appears to have been constructed by a series of compositionally similar granitic pulses. The absence of internal sharp contacts implies that this process may have been quite a rapid with some degree of homogenisation between these respective pulses, subsequent to their emplacement, now masking the original contacts.

The following section aims to investigate if the relationships seen in the Barnes Gap is laterally persistent when traced to the SW. In the areas of Losset and Leahanmore-Croaghacormick, some 2.5 km and 10.5 km to the SW of the Barnes Gap respectively, the granitoids (1-2 metres in length) were studied in an attempt to identify similarities with the field relationships seen in the Barnes Gap.

5:3:5 Losset

The pluton was studied in the vicinity of the junction between the Termon and Glenveagh-Churchill road (GR C 067221) (see figure 5:11).

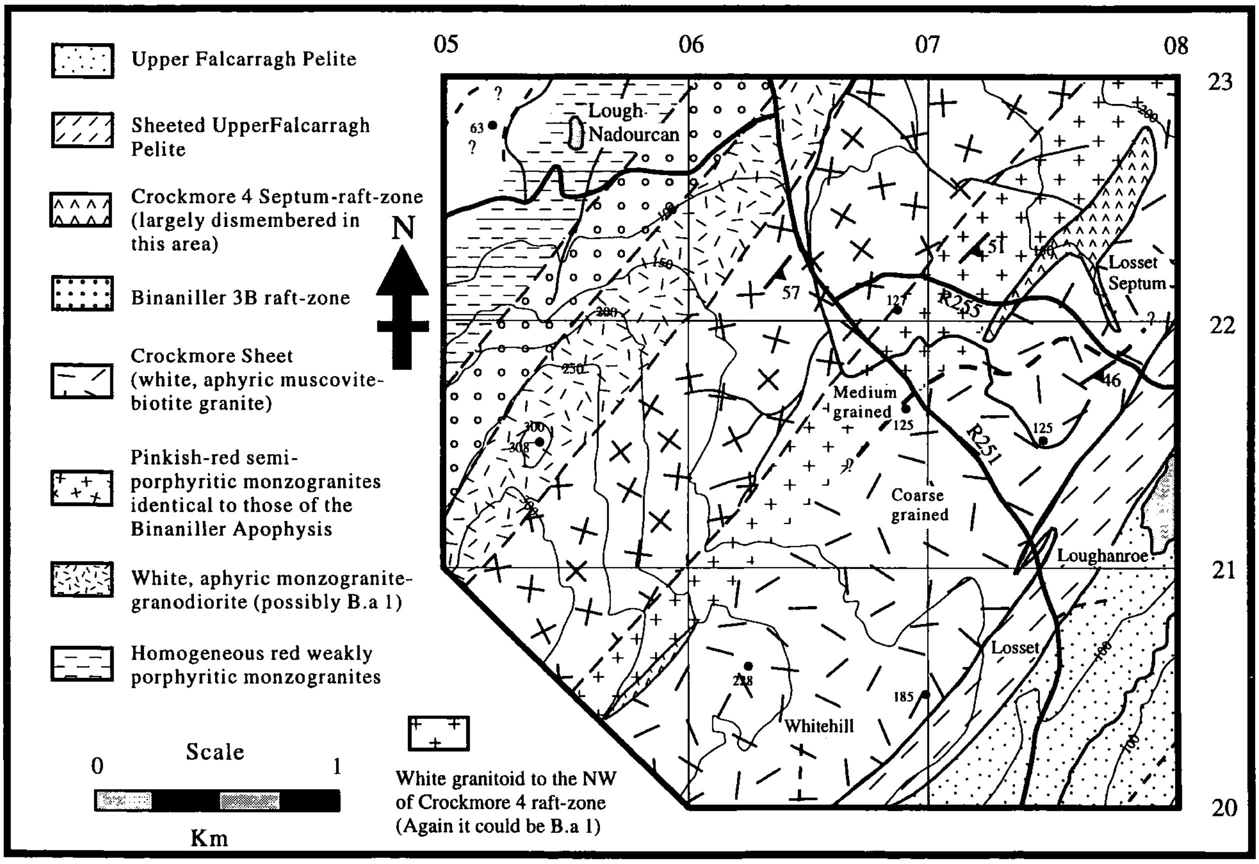


Figure 5:11:- The granitoids of the Losset area.

In the Barnes Gap it was observed that the Crockmore septum breaks up in to rafts 200 metres to the NE of the N56 main road. In the Losset area rafts of pelite and metadolerite, along strike from the Crockmore septum are clearly visible near the bend in the road (GR C 072221). In certain areas at Losset it appears that the septum has preserved some of its integrity and forms a prominent cliff, some 200m long, visible from the road some 300 metres to the NE of the above mentioned rafts (see figure 5:12). This septum (the Losset septum), dips very shallowly to the SE (035°/19°SE) with sheets of granitoid clearly intruded along the bedding or foliation within the semi-pelites (see figure 5:12).

Figure 5:12:- The Losset Septum shallowly inclined to the SE (Crockmore Sheet lies above it).

The granitoid above the semi-pelite septum is dominantly an homogeneous white biotite monzogranite identical to that seen in the Barnes Gap rail-cutting (figure 5:5), i.e. Crockmore Granite. In the Losset area the Crockmore-type monzogranite forms the main contact to the pluton and can be observed on the roadsides at (GR C 073211) and (GR C 079218). In both these areas the contact is quite sharp with the adjacent country rock pelites having been intruded by lenticular granitoid sheets, with the length of the sheets being proportional to the width, (see figure 6 Pitcher & Read (1959)). The composition of these sheets ranges from muscovite-garnet aplites to medium-grained granitoids similar to the Crockmore Granite in petrography. Within these marginal sheets deformation is intense with pervasively developed sinistral S-C' fabrics present with an associated stretching lineation plunging gently to the NE (see figure 5:13). As in the Barnes Gap section the deformation rapidly drops off within the Crockmore type monzogranite away from this contact.

To the NW of these pelitic rafts which form part of the dismembered Crockmore Septum (raft-zone 4) there is a white biotite-granitoid similar to the B.a. 1 monzogranite seen immediately to the NW of the Crockmore Septum at Barnes Gap.

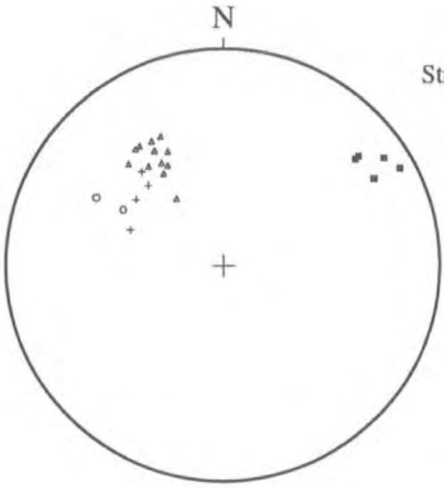


Figure 5:13
Structural data from the Losset area

This passes north-westwards into a coarse pink porphyritic monzogranite (200 metres SE of where the roads unite, GR C 069220) typical of the monzogranites in the Binaniller Apophysis. In the Losset area the outcrop width of these pink weakly porphyritic monzogranites is much less, about 400-500 metres wide. To the NW of these monzogranites there are more white feldspar, biotite granitoids which can be traced as far as the Binaniller 3B raft-zone. This raft-zone is well exposed in small roadside cuttings (GR C 058227) created by recent road straightening.

Figure 5:14:- White, equigranular biotite-rich sheets extensively sheeting the Binaniller 3B raft-zone (in greater density than in the Barnes Gap area).

Within these cuttings are extensive pelite rafts (no longer a septum as the rafts are clearly surrounded by granite) (see figure 5:14) which have been intruded by medium-grained, white, aphyric biotite granitoids that are identical in appearance to small sheets seen within the Binaniller Septum at Barnes Gap (i.e. figure5:8). To the NW of this 250 metre wide raft-zone are medium-grained reddish-pink weakly porphyritic monzogranites which are generally intense L-tectonites with the S_6 foliation only weakly developed and YZ surfaces of this granitoid type appearing almost undeformed. The absence of suitable strain markers prevents this strain from being quantified and its tectonic significance is not fully understood.

The granitoids to the SW of Barnes Gap and the granitoids in the Losset-Glenveagh area have been mapped by Pande (1954 *unpublished thesis*) as shown in figure 5:15. The mapping by the present author in the Barnes Gap and Losset agrees to some extent with the granitic facies identified by Pande (1954). The following list is correlated with the present authors scheme:-

i)Ga:- heterogeneous granitoids containing abundant Thorr Granodiorite xenoliths.

ii) Gd_2:- Crockmore sheet

iii) Gd_1:- this is the B.a. 1 white biotite granitoid to the immediate NW of the Crockmore Septum.

iv) Gb_3:- this is the pink porphyritic monzogranites of the Binaniller apophysis, although Pande (1954) does not differentiate the grain size variations.

v) Gc:-white porphyritic granitoids. The present author did not identify this facies in the Barnes Gap area, although "white" more equigranular granitoids were encountered within the pink porphyritic monzogranites of the Binaniller apophysis. In the Losset area these white granitoids were seen in the near vicinity of the Binaniller 3B raft-zone.

vi) Ge_1:- these are the heterogeneous granitoids containing xenoliths of quartzite and semi-pelite.

5:3:6 Leahanmore - Croaghacormick

In this area a search for the granitoids resembling the Crockmore and Binaniller apophyses was made. Figure 5:16 shows the area of study. This area is most easily accessed from the track to the NW of Gartan Lough, some 2.25 km to the SE of Leahanmore. In the Losset area the Crockmore-type monzogranite forms the main margins to the Main Donegal pluton. In the Leahanmore-Craoaghcormick area the relationship is complicated by large amounts of pegmatitic material. The pluton contact is located 50-100 metres to the NW of the deer fence (GR C 036152) where the country-rock is Lower Falcarragh Pelite (Pitcher & Shackleton 1966). Also seen along this contact zone was abundant exposures of mafic-rich granitoid which resembles that of the Thorr pluton. The significance of these exposures will be addressed in the following chapter. The marginal part of the pluton consists of

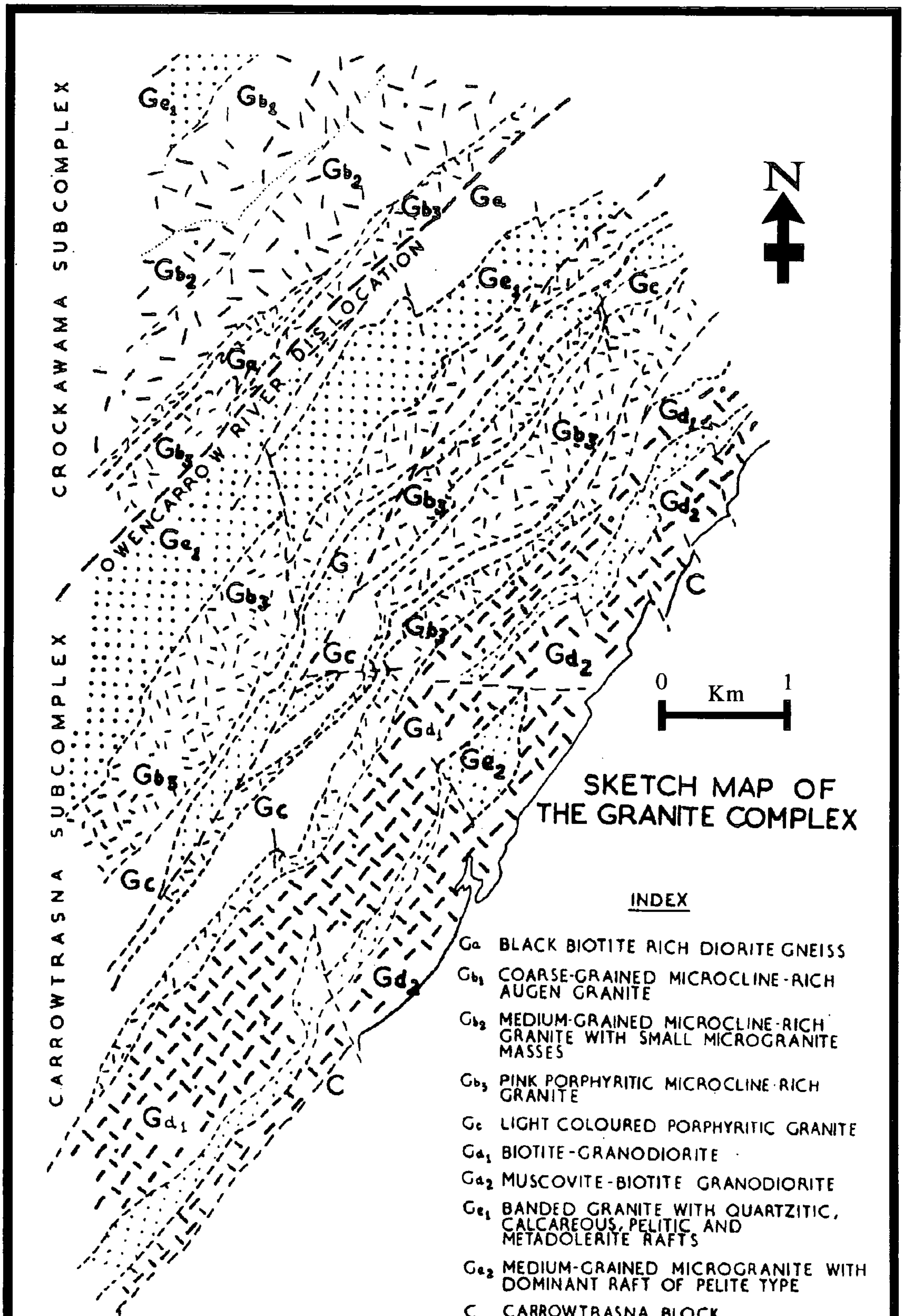


Figure 5:15:- The granitoids of the Stragraddy-Losset area (*after* Pande 1954)

medium-grained, aphyric biotite granitoids which form a crude banding caused by fluctuating biotite content and grain size. Also present are coarse, large irregular masses of pegmatite. Further raft-zones of Thorr Granodiorite xenoliths up to 100 metres wide are encountered within the two above mentioned granitoid types.

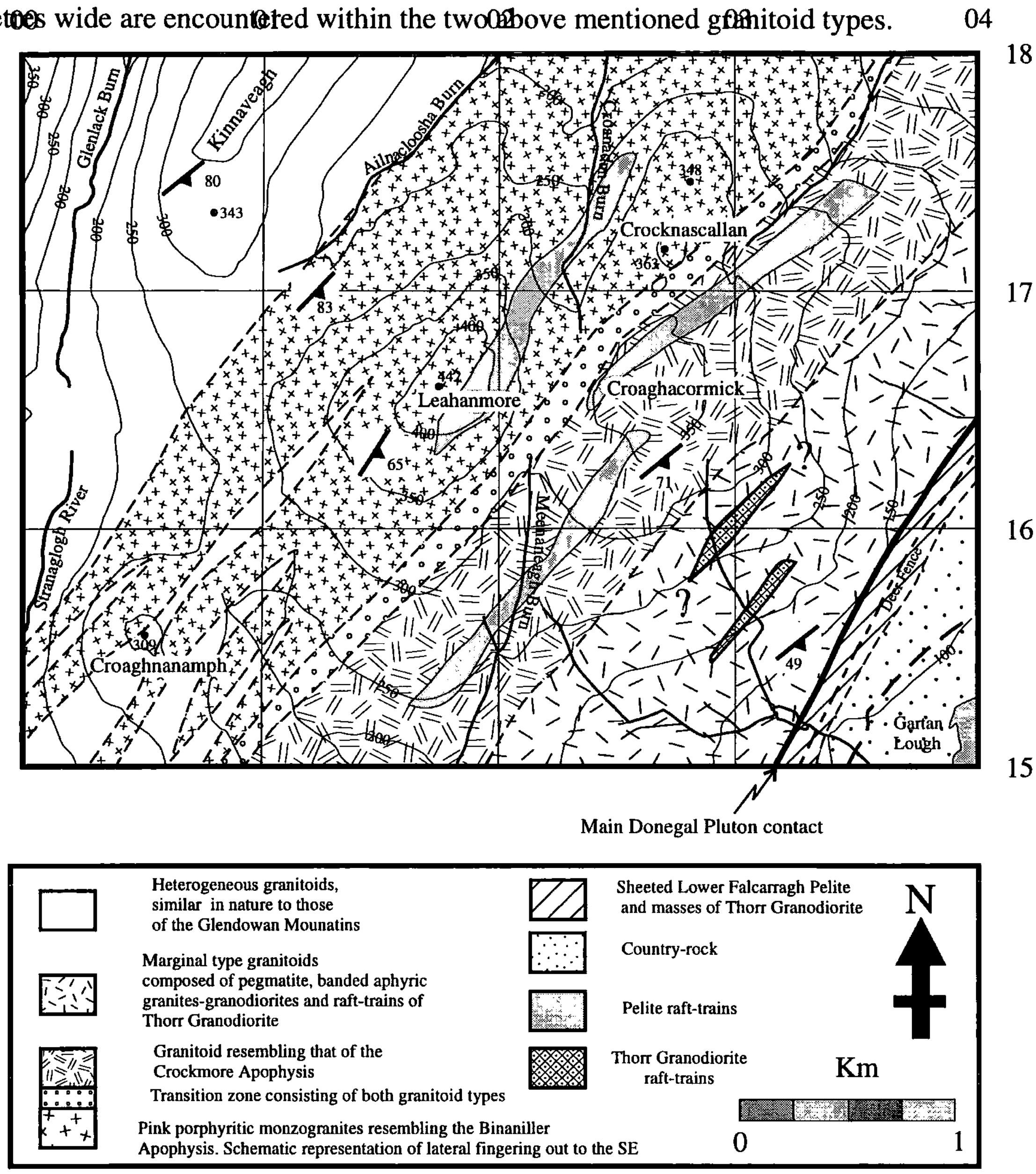


Figure 5:16:- The granitoids of the Leahanmore-Croaghacormick area.

Approaching the summit of Croaghacormick from the SE, granitic material resembling that seen in the Crockmore apophysis is encountered at GR C 028161 and is accompanied by a notable decrease in the amount of pegmatitic material. Although the majority of the pegmatite decreases there is some pegmatitic material within

Crockmore monzogranite, a feature not observed in the Barnes Gap and Losset areas further to the NE.

Rafts of pelite and metadolerite become quite abundant in the Leahanmore - Croaghacormick area forming three main swarms which are believed to belong to the dismembered Crockmore Septum (Pitcher & Read 1959). The orientation of the rafts is quite variable in this area, ranging from being vertical to dipping shallowly to the SE. These rafts are relatively unaltered with the planar schistosity still preserved. The width of this "white", equigranular, biotite monzogranite zone is up to 750 metres extending from the summit region of Croaghacormick to the SE slopes of Leahanmore (GR C 017166).

At the watershed between Croaghacormick and Leahanmore (GR C 023165) there is a transition zone where the granitoids are more heterogeneous consisting of bands of the Crockmore type facies together with a medium-grained pink porphyritic monzogranite. The latter of these granitoids resembles that of the Binaniller apophysis, although in this area they are generally more coarser. The coarser nature may be due to greater heat insulation, i.e. deeper within the pluton and hence promoting larger grain growth. The width of this transition zone is 50-75 metres and to the NW it passes into relatively homogeneous, coarse, pink porphyritic monzogranites that can clearly be observed in the summit region of Leahanmore. On the NW slopes of Leahanmore the grain size becomes finer within the pink porphyritic monzogranites, with occasional zones of white biotite granitoid becoming more common, notably towards the watershed between Leahanmore and Mount Kinnaveagh. These white, biotite granitoids were observed in the Losset area close to the Binaniller 3B raft-zone and at Barnes Gap (immediately above this same raft-zone). The above mentioned watershed lies along strike from the Binaniller 3B raft-zone, although no pelitic material was observed in this area perhaps due to very poor exposure. On Kinnaveagh the granitoids are much more heterogeneous with fine-grained granodiorites and medium-grained tonalites forming pod-like masses in the banded weakly porphyritic white monzogranites which resemble GM 4 of the Glendowan Mountains (chapter 4; section 4:4). The style of emplacement on Kinnaveagh is strongly reminiscent of that seen in the Croaghacullin area of the Glendowan Mountains and it is highly probable that the Kinnaveagh area represents an extension of this heterogeneous zone (i.e. the "Glendowan Mountains package"). It is therefore the authors belief that the Binaniller 3B raft-zone may sub-crop in the Leahanmore-Kinnaveagh area as its projected location does separate different granitoid types, i.e. the relatively homogeneous monzogranites of Leahanmore from the more variable granitoids of Mount Kinnaveagh.

The degree of deformation within these granitoids is relatively high especially within the facies that resembles the Crockmore apophysis. Despite the S_6 foliation being present (see figure 5:17) the granitoids show a much stronger lineation (emphasised by a strong preferred alignment of biotite). As with the Losset area, the reason for these higher prolate strains in this area is not fully understood. Towards Mount Kinnaveagh this component of constriction diminishes accompanied with an increase in the overall inclination of the S_6 foliation.

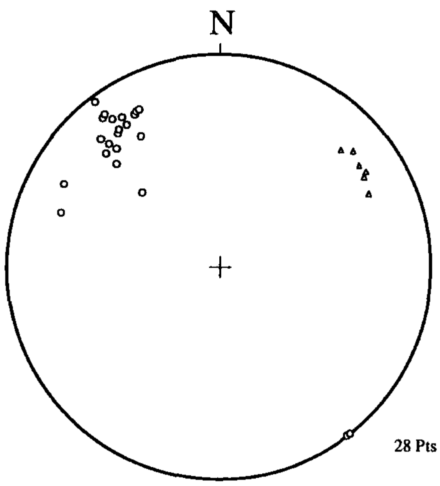


Figure 5:17

S6 foliation (O) and associated stretching lineation (△) in the Leahanmore-Croaghacormick area.

In summary the granitoids of Croaghacormick and Leahanmore do show similarities with the Crockmore and Binaniller apophyses seen in the Barnes Gap, although the relationship in this former area is by no means as clear-cut as that observed in the latter. The following list describes some of the main differences:-

1) In the Barnes Gap and Losset area the Crockmore apophyses formed the immediate contact of the pluton and contained relatively little pegmatite. In the Croaghacormick area the homogeneous Crockmore apophysis is not encountered for approximately 800-900 metres to the NW of the main pluton contact. These intervening granitoids consist of large volumes of pegmatite, banded biotite granitoids and several raft-trains of Thorr Granodiorite. These variable granitoids are typical of the marginal facies and are commonly seen along the south-eastern margin, to the SW of the Lough Gartan area, and along the entire length of the NW margin of the Main Donegal Pluton (see section 5:3 for more details).
2) The rafts which lie along strike from the Crockmore Septum become progressively wider spaced with the exact transition between the Crockmore and Binaniller apophyses becoming much less obvious and often being transitional in nature. Furthermore the Binaniller 3B raft-zone is not present in this area although its predicted location does approximately form the NW boundary of the

relatively homogeneous Binaniller pink porphyritic monzogranites with the more heterogeneous granitoids to the NW.

3) The lateral persistence of the Crockmore and Binaniller apophyses to the SE of this area is uncertain as the exposure is relatively poor. In the Croaghacullin area no granitoids resembling the Crockmore sheet were observed implying this apophysis may finger out into heterogeneous granitoids within the intervening ground.

In the Losset area and particularly the Barnes Gap, the Binaniller apophysis was typified by its overall homogeneity, apart from grain-size variations, and absence of any inclusions of country-rock or older granitoids. On Leahanmore this apophysis is not as homogeneous with inclusions of older granitoids becoming common. In the summit region (GR C017166) there are darker bands of biotite-rich tonalite aligned parallel to the foliation. The appearance of these bands resembles the GM 2 tonalite seen in Map I of the Croaghacullin area. On Croaghnanamph, 1 km to the SW of Leahanmore the pink porphyritic monzogranites are still abundant, but other granitoid facies of the Glendowan Mountains area are becoming more common. Following this train of thought the pink porphyritic monzogranite (GM 5) seen in Map I to the SW of Croaghacullin shows a striking resemblance to the coarse, pink porphyritic monzogranites exposed in the summit region of Leahanmore. The absence of continual exposure in the Bullaba Valley prevents this correlation from being unequivocally proven, however the evidence suggests the Binaniller apophysis fingers out laterally as it intrudes, in a south-westerly direction, into the older more heterogeneous zones of the pluton (see figure 4:1 or Map D of this thesis).

In the Barnes Gap at the NW end of the pluton it appears that both the Binaniller and Crockmore apophyses initially intruded into meta-sediments as is clearly indicated by the relationships between these granitoids and the dominantly autochthonous septa which form their boundaries. In a south-westwards direction these apophyses appear to transgress across the former contact of an earlier, more granodioritic and tonalitic pluton, which existed in this area prior to the emplacement of the these two relatively homogeneous masses of monzogranite. The significance of this possibly earlier contact will be discussed in chapter 8.

5:3 The granitoids of the NW region of the pluton

The country rock along the NW margin is dominantly Thorr Granodiorite, although to the NE of Cashel Mountain (SE of Creeslough) the Ards Quartzite forms the immediate envelope with the Thorr Granodiorite now occurring as xenoliths within the Main Donegal Granite. The north-western regions of the pluton are typified by large areas of very homogeneous, dominantly pinkish-red porphyritic monzogranites. These monzogranites are typically seen to the NW of the Derryveagh 2 raft-zone, a consistent relationship seen at several locations along the length of the pluton (see figure 4:1 or Map D). To the NW of these homogeneous pink porphyritic monzogranites, along the marginal regions of the main NW countryrock contact, the granitoid facies are typically more heterogeneous with variable sheets containing xenoliths of the country-rock common together with pegmatite and microgranite dykes plus a series of banded aphyric, medium-grained granitoids. One must emphasise that within this zone the integrity of the countryrocks is lost; with it being entirely xenolithic. This may be the result of extremely dense granitoid sheeting. The country-rock to xenolithic marginal facies transition is well seen along the entire length of the NW margin with the width of the marginal facies varying from 175 metres to up to 700 metres. It is particularly well seen at the NW entrance of the Poisoned Glen and along the Lackagh River, in the vicinity and especially to the SE of the bridge. Pitcher & Read (1959) called this xenolithic margin the "*Border 1 raft-zone*". The present author favours the term "*marginal facies*" (marginal granitoid facies package) as the rafts which form this zone do not have the typical appearance of the raft-zones seen within the pluton.

These granitoids in this north-western part of the Main Donegal Granite have been studied at a number of locations; the Lackagh Bridge-Cocks Heath Hill area, the Kingarrow area, the Dunlewy-Poisoned Glen area and at the Brockagh-Meenderryherk area. The most detailed area of study was at Lackagh Bridge and thus will be described first. In each of the areas the country-rock and *marginal facies* will be discussed first with the homogeneous monzogranites being described after.

5:3:1 Lackagh Bridge

In this area the marginal granitoids and homogeneous pink porphyritic monzogranites were studied along the Lackagh River and also on the well exposed glacial surfaces to either side of the Glen-Lackagh Bridge road (see figure 5:18).

5:3:1:1 The Marginal facies

The salient features of the marginal zone of granitoids is shown in figure 5:19 in a section along the Lackagh River, immediately to the SE of the road bridge (GR C 095310), which crosses the contact of the Main Donegal Granite. The contact occurs where sheeted autochthonous country-rock becomes xenolithic within variable granitoids, (i.e. pegmatite and banded monzogranite to granodiorite). This rapid transition from sheeted country rock to xenolithic country rock is probably related to the increasing abundance of sheets towards the main pluton body causing the sheeted zone to break-up.

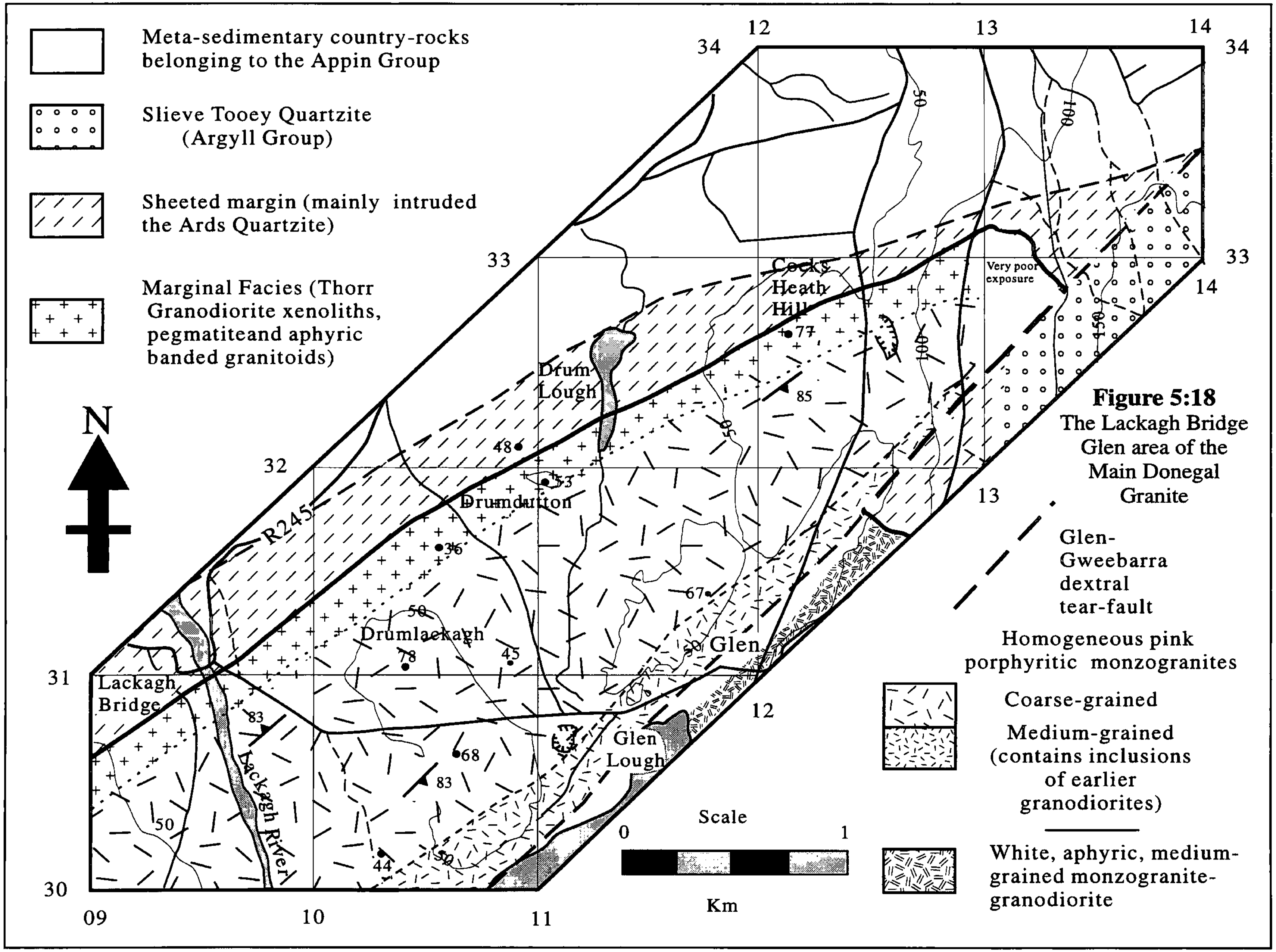
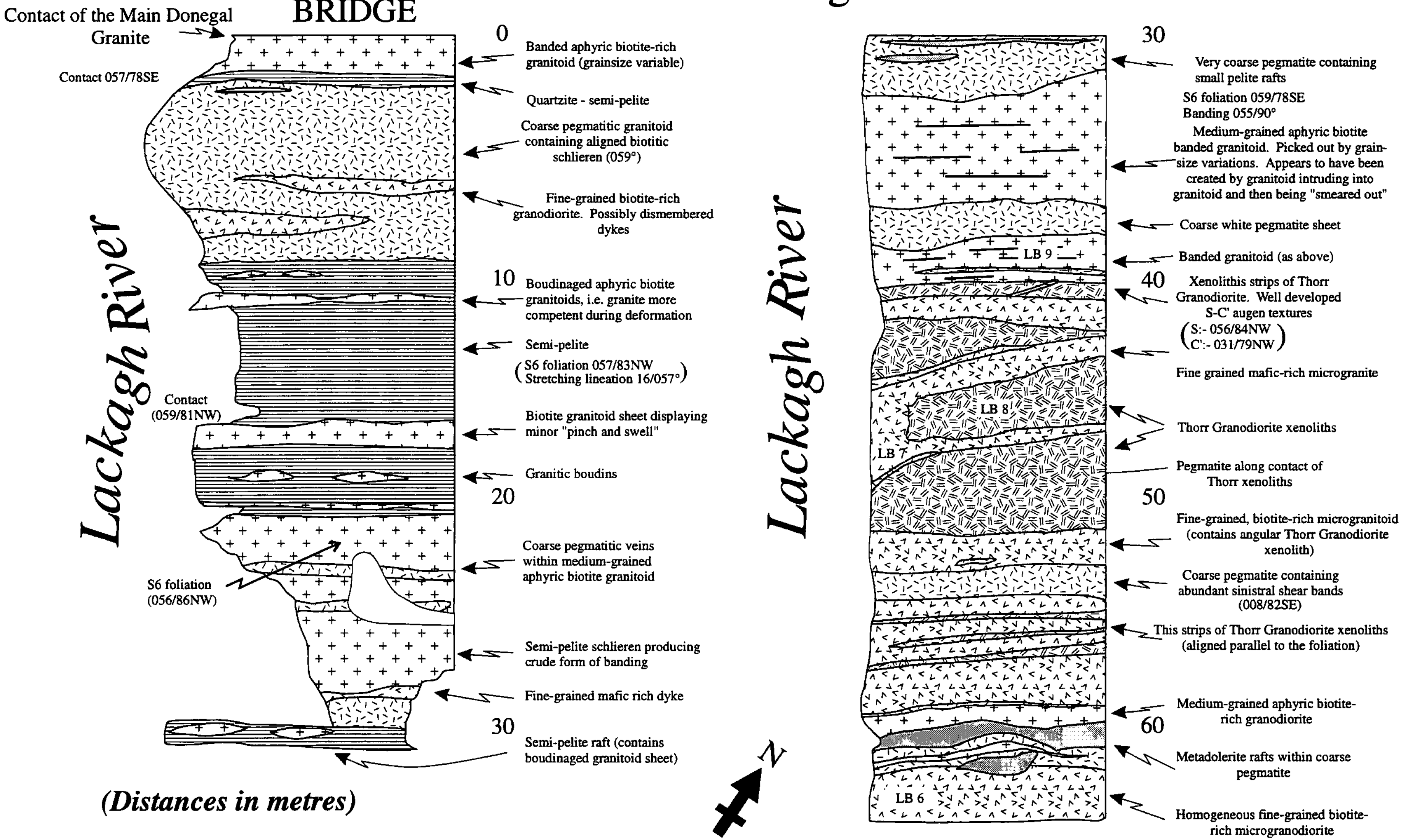


Figure 5:18
The Lackagh Bridge Glen area of the Main Donegal Granite

The contact of the pluton lies underneath the bridge, as the rocks exposed in the small roadside quarry immediately to the NW of the bridge is clearly sheeted Ards Quartzite. Along the section in figure 5:19 there are rafts of Thorr Granodiorite, metadolerite and pelite and semi-pelite occurring within younger granitoids of the Main Donegal pluton. No autochthonous Thorr Granodiorite (TG) was seen in this area with the immediate country-rock being Ards quartzite. The oldest of these marginal granitoids is a medium-grained aphyric biotite granitoid which is commonly banded parallel to the foliation within the pluton. Banding is demarcated by grainsize variation and variable proportions of biotite. This banded granitoid strongly resembles the more mafic finer grained granodiorites and tonalites seen in the central portions of the pluton, i.e. SRU 1 and SRU 2 of chapter 4. These banded granitoids range from 1-8 metres in width and tend to lie along the contacts of the country-rock rafts, i.e. when younger pegmatites and fine-grained monzogranite dykes are "removed" this granitoid has sheeted geometries with the country-rock. The banding is interpreted to be the result of deformation of a crystallising mush around more competent rafts with the "spalling off" of smaller fragments from larger rafts and subsequent streaking out of this material enhancing the banding. The next phase of intrusion in this marginal zone is a suite of coarse pegmatites which tend to intrude into semi-pelitic material or into the banded granitoids already described. The very coarse nature of this rock promotes the development of large sinistral shear bands, oriented at ~020°. The youngest phase within the marginal facies of this area is a fine-grained granodiorite ($\leq$2 mm) which occurs as small dykes which intrude all other lithologies along the section (apart from one late pegmatite sheet at 53 metres (figure 5:19) which intrudes it). These fine-grained dykes only become abundant at 40 metres from the pluton contact where they dominantly intrude into xenolithic masses of TG. This fine-grained granodiorite usually forms sheets, usually less than 2 metres wide, which are common at many locations along the NW margin of the pluton. Between 50-60 metres from the contact the break-up of Thorr Granodiorite rafts, by the dykes, into strips also contributes to the banding in this marginal phase. Despite these dykes being most common in this marginal facies they are also seen, to a lesser extent, intruding into the more homogeneous coarse pink porphyritic monzogranites to the SE: a relationship which is very well displayed in Cocks Heath Hill Quarry, to the NE of Lackagh Bridge (see section 6:2:1:2 of Chap. 6).

The width of the marginal facies in the Lackagh Bridge area is between 180-200 metres whilst on Drumlackagh Hill to the NE it is up to 225 metres wide. To the SE of this zone one encounters very homogeneous coarse pink porphyritic monzogranites which show only very minor variation over a large area. The boundary between the marginal facies and these monzogranites is transitional over a distance of

Figure 5:19:- Section along the Lackagh River across the contact of the Main Donegal Granite



50-75 metres where the thin bands of porphyritic granite become thicker and are reciprocated by a decrease in width of the marginal facies bands until eventually the porphyritic bands coalesce in to a homogeneous mass. In general this transition is best exposed on the NE side of the Lackagh River at GR C 09630308.

5:3:1:2 The homogeneous pink porphyritic monzogranites

The pink porphyritic monzogranites of this area are easily observable on either side of the Lackagh Bridge-Glen road, notably at the bend in the road (GR C 100307) where a small track extends SE across well exposed glacial pavement. Weathered exposures have a distinct porphyritic appearance with K-feldspars up to 8-10 mm in diameter. Fresh surfaces reveal that these megacrysts have distinct reddish-pink colour. The groundmass is generally much finer grained (1-3 mm) consisting of white plagioclase, biotite and quartz (see figure 5:20). The porphyritic nature of this monzogranite leads to the excellent preservation of S-C' fabrics. Although the shear sense is sinistral, the less common antithetic set is also clearly developed (see also figure 5:23a). These pink porphyritic monzogranites extend south-eastwards to the shores of Glen Lough, although they become highly weathered and fractured due to the presence of the Glen-Gweebarra dextral transcurrent fault on the NW side of the Lough.

Figure 5:20:- pink porphyritic monzogranite from Glen quarry (GR C125326). Note strong S-C' fabrics implying sinistral shear. Antithetic C' plane also developed.

Despite the overall homogeneity of these porphyritic monzogranites, very minor grain-size variation is encountered. On the whole contacts tend to be transitional. Mapping of the monzogranites in the Lackagh Bridge area demonstrates that this variability is more common in the SE part, in the ground to the immediate NW of Glen Lough. At GR C 108303 there are well exposed glaciated surfaces which display zones of weak banding. The zones of banding strike parallel to the S_6 foliation and consist of pink porphyritic banded monzogranite with medium-grained aphyric biotite-rich granitoid. This latter granitoid type strongly resembles the banded biotite-rich facies observed within the marginal zone along the NW margin of the pluton. In the granite quarry along the Glen-Lackagh Bridge road (GR C 112307) there are two varieties of porphyritic monzogranite although no distinct contacts were observed despite the good quality of exposure. The variation within these pink porphyritic monzogranites was also been observed by Pande (1954) along strike to the SW of the N56 road (see figure 5:15). This author also states that in the NW there are coarse-grained pink porphyritic monzogranites which change by transition into a very slightly finer grained porphyritic monzogranite with the latter containing xenoliths of highly altered pelite and limestone material. Although the present author did not observe xenoliths of meta-sedimentary material he agrees that the south-eastern outcrop of the pink porphyritic monzogranites is typified by the more common occurrence of xenolithic material. In the Lackagh Bridge area these xenoliths appear to be older granitoids which may belong to the marginal facies. Furthermore in the map of Pande (1954) there is the marginal facies along the NW margin of the pluton (Ge_1) which he described as essentially banded granite of varying composition in which xenoliths of country-rock are observed.

5:3:1:3 Deformation in the Lackagh Bridge area

The strain within these granitoids is very high within this area with well developed S-C' fabrics seen within all the different facies observed. The S_6 foliation in this area is generally uniform (~055-060°) and is steeply inclined to the SE (see figure 5:21). The coarse porphyritic monzogranites and Thorr xenoliths (see figure 5:22) are generally converted to augen gneisses by the deformation with well developed S-C' fabrics. The sense of shear is dominantly sinistral although there is a numerically subordinate antithetic set. Hutton (1977) documented that these structures are common both to the pluton and the adjacent envelope rocks implying the "granite" is being deformed by an external tectonic force, i.e. a sinistral shear zone. Granite dykes have been boudinaged to varying degrees depending upon the competency contrast between the granite and its host. Therefore dykes are most intensely boudinaged where sheets are intruded within pelites, although larger, thicker sheets only show minor “pinch and swell”. Pegmatites or microgranite dykes intruded into granite show little sign of boudinage implying very similar ductility

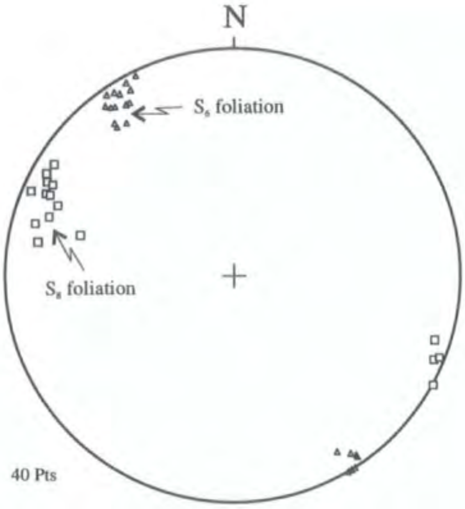


Figure 5:21
Stereogram of the S_6 and S_8 foliation within the marginal granitoids and the pink porphyritic monzogranites within the Lackagh Bridge area.

during deformation. Pitcher & Berger (1972) observed similar phenomena in the Glenleheen area of the pluton where late microgranites appeared undeformed in the granitic host, but where the same dykes crossed a limestone raft they became intensely deformed. The presence of boudinaged veins demonstrates that deformation continued until the granitoids sheets were well below the RCMP with these sheets tending to behave more competently than the host rock. To what late stage this deformation continued will be addressed in the next section.

Figure 5:22:- hand specimen of Thorr Granodiorite from Lackagh River section (fig. 5:19: sample LB 8) showing S-C "augen" textures. High strains around feldspar poprhyroclasts result in strongly flattened quartz grains around their margins.

Thin-Section

Within the marginal granitoid facies in the Lackagh Bridge area thin sections of the relatively coarse, pink porphyritic monzogranite and the mafic-rich, fine-grained granodiorite dyke were examined to compare deformation styles (see figure 5:23 a & b respectively). In outcrop the porphyritic monzogranite resembles augen gneiss due to the presence of larger microclines. Both feldspars show strong evidence of mechanical distortion, although no microfracturing was observed within these porphyroclasts. Twin lamellae in plagioclase and microcline are commonly distorted with crystals showing sweeping undulose extinction. This may be caused by a combination of dislocation glide or sub-microscopic brittle fracturing within these grains. Along the margins of the feldspars, where strain is very high, there is athin rind of recrystallised feldspar forming "core and mantle" structures (generally finer-grained than recrystallised quartz as shown in figure 5:23 c). The uniform grain size and very fine-grained nature argues against cataclasis within these grains and favours dynamic recrystallisation. In some microclines there are discrete fractures filled with fine grained feldspar. This structure may be relic sub-magmatic deformation as described by Bouchez *et al.* (1992). Microstructural evidence of plagioclase indicates that is has deformed at temperatures between 350-450°C. Quartz shows considerable evidence of grain size reduction and varying degrees of undulose extinction. The highest strained quartz tends to be flattened parallel to the foliation and contain well developed subgrains. Larger quartz porphyroclasts show quite strong undulose extinction whilst the boundaries of these crystals have well developed subgrains and highly irregular lobate boundaries indicating extensive dynamic recrystallisation by SR as well as GBM. This leads to the development of relatively small quartz grains which appear strain free. Biotite is dominantly wrapped around feldspar porphyroclasts producing "eye" like structures.

Within this sample there is no evidence for overall brittle deformation as quartz behaved in a totally ductile manner. Feldspar is showing a combination of brittle microfracturing and some degree of recrystallisation along feldspar margins. Generally these fabrics indicate high-temperature deformation (temperatures equivalent to upper greenschist facies).

In the fine grained microgranodiorite (figure 5:23b); there are well developed S-C' fabrics with the distribution of strain being more uniform due to its equigranular nature. Plagioclase again shows no evidence of brittle microfracturing and tends to be undeformed apart from the presence in some crystals of bent twins. Fine-grained recrystallised feldspar was only seen along a number of plagioclase boundaries. Overall there is a weak alignment of feldspar laths parallel to the "S" foliation (S_6 of

Hutton 1982). Quartz has undergone extensive recrystallisation tending to be quite fine-grained, although larger, more highly strained quartz porphyroclasts still remain.

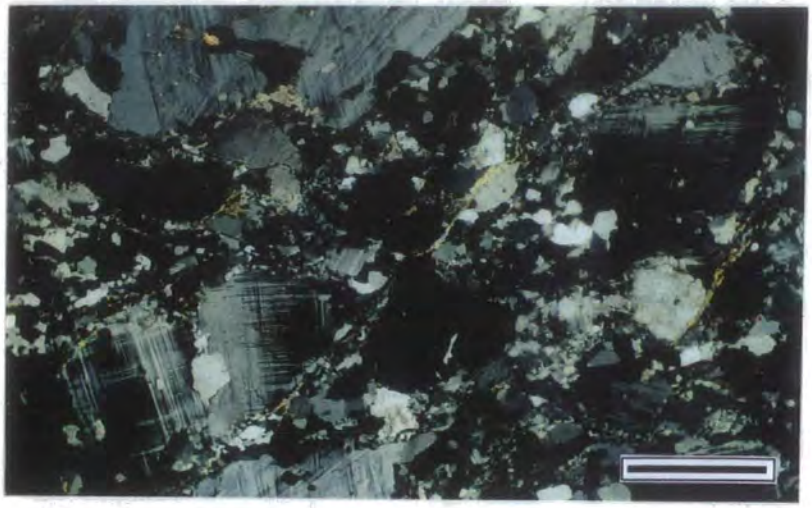


Figure 5:23a
Photomicrograph of the pink porphyritic monzo-granite. (scale bar = 1mm)

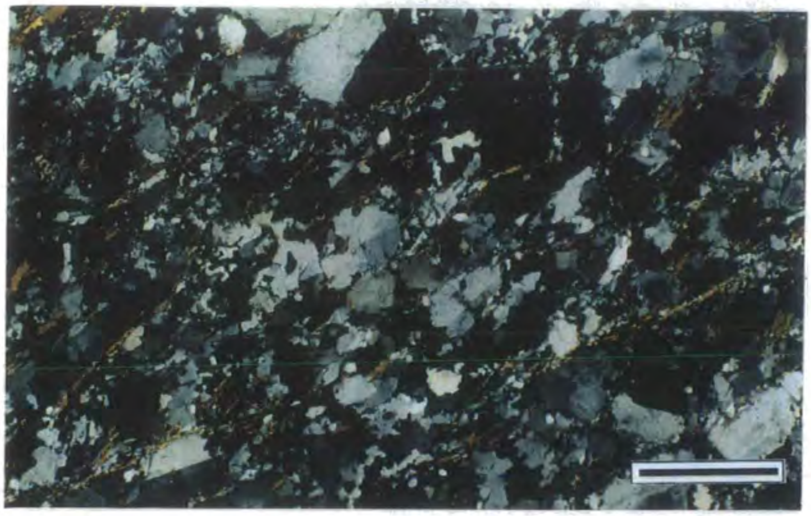


Figure 5:23b
Photomicrograph of fine-grained granodiorite dyke seen in the marginal facies. (scale bar = 0.5 mm) (see text).

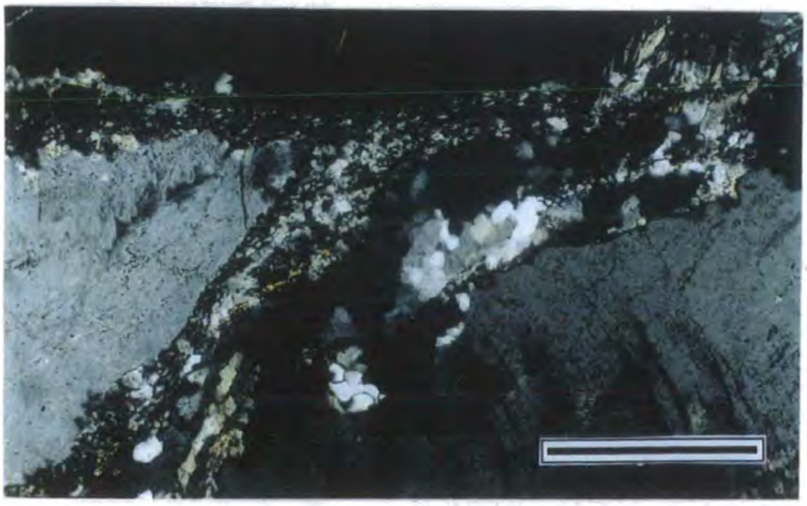


Figure 5:23c
Dynamic recrystall-isation in both quartz and feldpar forming fine-grained aggregates (K-feldspar is the very fine-grained material). (scale bar = 0.25mm)

The remainder of this section will address the lateral continuity of the pink porphyritic monzogranites to the SE of Lackagh Bridge, i.e. at Kingarrow, Poisoned Glen and at Brockagh-Meenderryherk.

5:3:2 Kingarrow

In this area the Thorr Granodiorite "prolongation" forms the immediate envelope to the Main Donegal Granite (see figure 5:24) and contains abundant MDG sheets of varying composition (trending 056°).

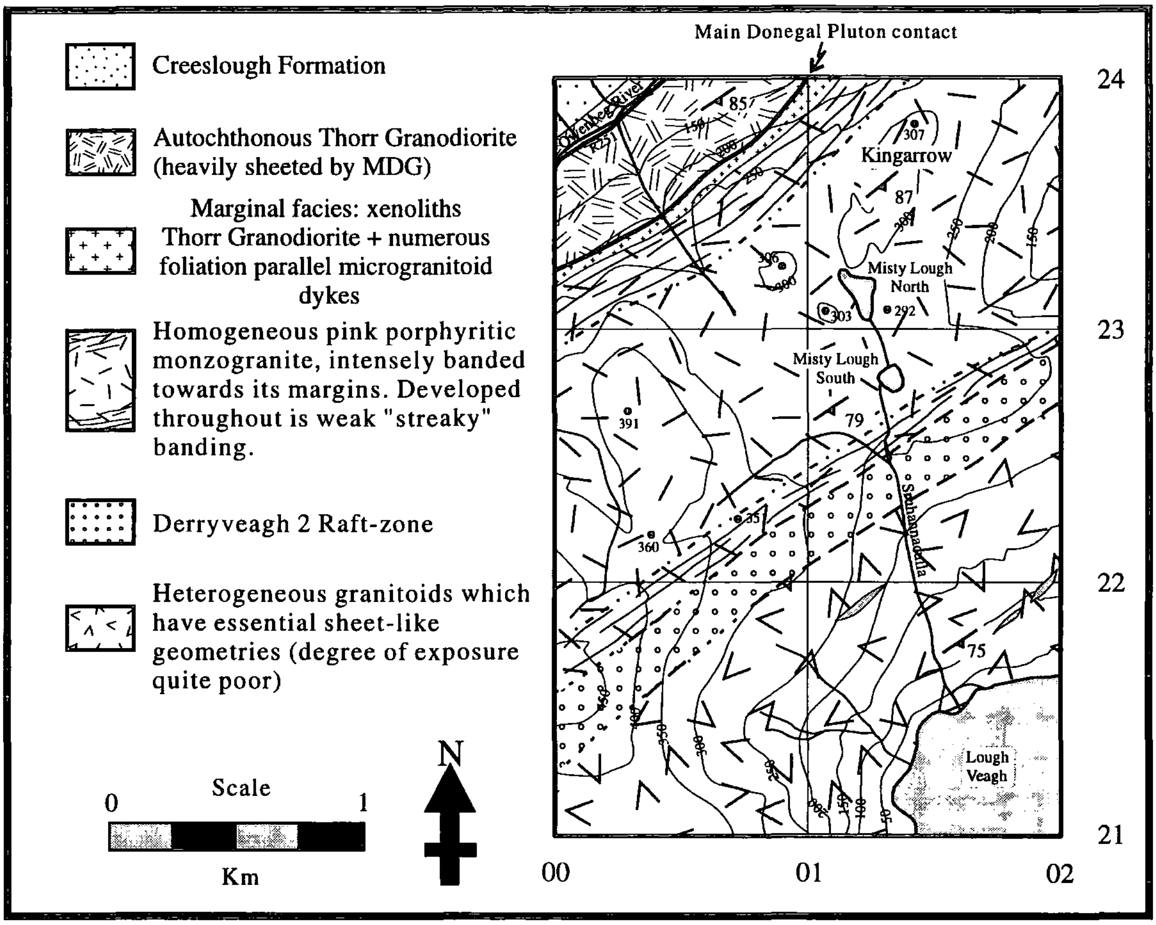


Figure 5:24:- The granitoids of the Kingarrow area, Derryveagh Mountains.

In this area the Main Donegal Pluton starts to form prominent topography with the external contact zone situated along a steep slope. The contact zone is easy to identify mainly due to the sudden occurrence of Thorr xenoliths rather than autochthonous TG. The marginal zone in this area is also marked by the occurrence of banded granitoid striking parallel to the NE-SW foliation. The banding is due to:- the intrusion of fine-grained, biotite-rich microgranodiorites; the streaking-out of TG xenoliths and the presence of compositionally dissimilar granitoids (a type of regular banding). In this marginal zone pink porphyritic monzogranites are common and are highly deformed, having an almost augen appearance. The sense of shear, as indicated by the C' shear band cleavage, is consistently sinistral. This zone of intense banding is confined to within an area 250 metres SE of the Main Donegal Granite contact. These pink porphyritic monzogranites are almost identical in appearance to those described at the Lackagh Bridge area with the exception that banding, on various scales, is much more common. The banding has a streaky appearance and is mainly the result of more biotite-rich monzogranite. These banded monzogranites outcrop for a width of ~1 km and are visible as far south as Misty Lough South (GR C 013228). Misty Lough South is situated within a prominent gully which is controlled by a late stage NE-SW fault (clearly visible on aerial photographs). To the NE of this area the fault slowly swings south-eastwards and eventually joins with the Glen-Gweebarra wrench fault in the Newbridge area (NW end of Barnes Gap). To the SE of Misty Lough South banding within the pink porphyritic monzogranites rapidly intensifies to form a regular-streaky type banding. This banding prefaces the presence of the nearby Derryveagh 2 raft-zone which outcrops 125-150 metres to the SE of the small lough. In this area the Derryveagh 2 raft-zone is 250 metres wide and consists mainly of Thorr Granodiorite, pelite, calc-pelite and quartzite. As already mentioned in chapter 4 the granitoids to the SE of this prominent raft-zone are finer grained and generally more heterogeneous.

In summary the granitoids of the Kingarrow area are homogeneous to the NW of the Derryveagh 2 raft-zone and consist of pink porphyritic monzogranites which in places are banded. Due to the relatively poor quality of exposure the variation within this monzogranite, as seen in Lackagh Bridge by the present author and by Pande (1954) in the area south of Creeslough, was not observed in the Kingarrow area.

5:3:3 Dunlewy - Poisoned Glen

In this area the marginal granitoids and pink porphyritic monzogranites were studied in a transect from the Dunlewy area extending southwards along the well exposed SE side of the Poisoned Glen up to the point where the Derryveagh 2 raft-zone is encountered which occupies the high ground which forms the watershed between the north-westerly flowing Croaniv Burn (flows through the Poisoned Glen) and the south-easterly flowing Sruhanavarnis valley. The distribution of the granitoids are shown in figure 5:25. The width of this marginal facies in the Poisoned Glen area is between 650-700 metres and is composed of abundant TG xenoliths, two phases of pegmatite, meta-sedimentary material and biotite-rich, fine to medium-grained granitoids. To the SE of these are homogeneous pink porphyritic monzogranites which are up to 1.6 km in width in this area. The following section describes the marginal facies and the porphyritic monzogranites observed along a section from the NW to the SE.

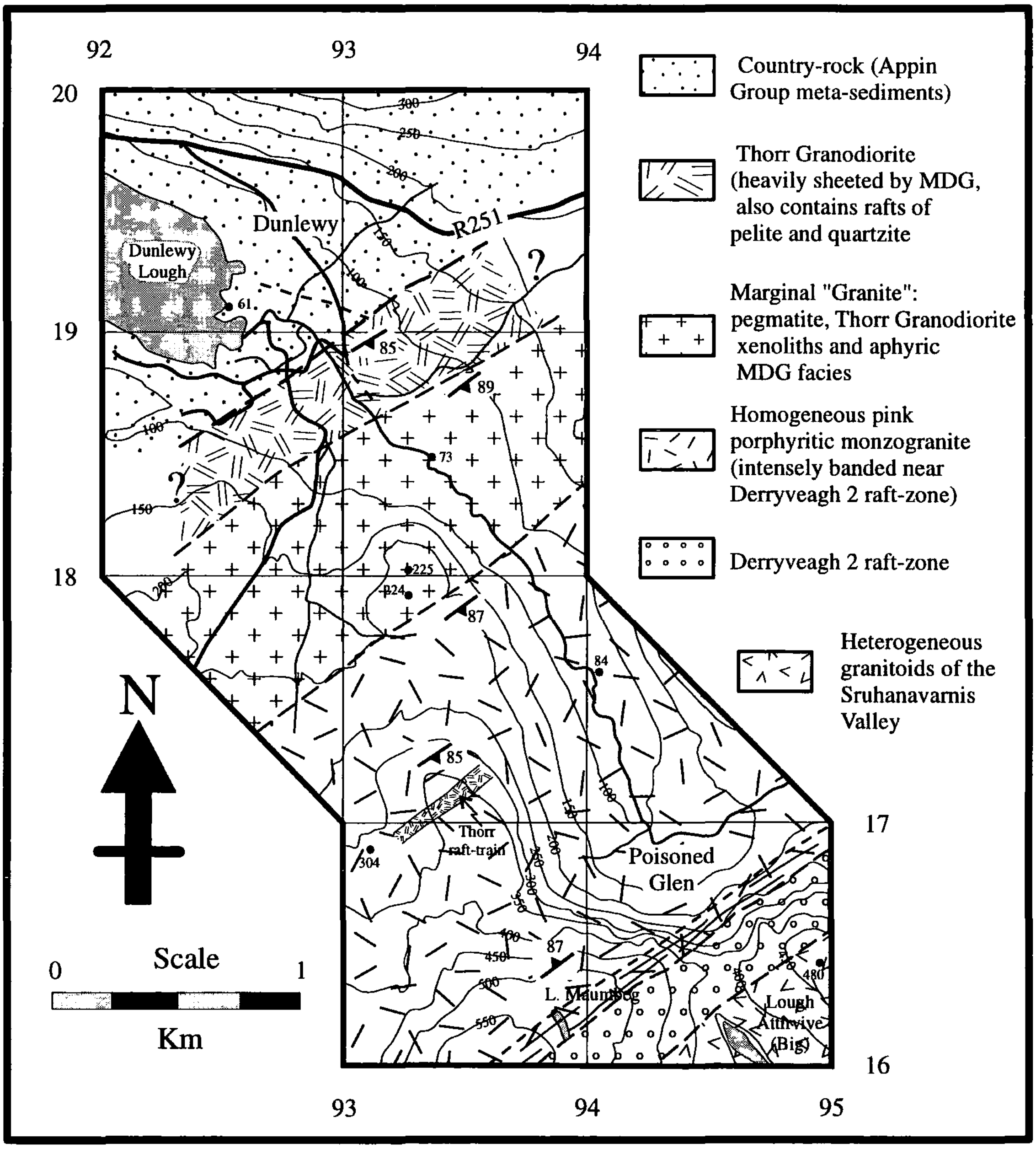


Figure 5:25:- the granitoids of the Dunlewy-Poisoned Glen area.

5:3:3:1 The contact zone and the marginal facies

The section begins near the old church at Dunlewy (GR B 930191) where the immediate envelope to the Main Donegal Granite in this area is the autochthonous strip of Thorr Granodiorite (the "prolongation" of the Thorr Pluton) which has been intruded by small granitoid sheets belonging to the Main Donegal Granite. The envelope to the Thorr Granodiorite prolongation is pelite and metadolerite belonging to the Creeslough Formation (Hutton & Alsop 1995). The contact of the Thorr Granodiorite with the country-rock is quite sharp and is clearly visible at GR B 930190. At this location there is a strong foliation (steeply inclined to vertical) whose orientation is similar to that of S_6 (of Hutton 1982) although no S_8 (C' foliation) was visible. The mineral stretching lineation in the Thorr Granodiorite, MDG sheets and meta-sediments plunges shallowly to the SW. In the following 200 metres SE of this location the S_8 cleavage develops and intensifies and indicates sinistral shear (see figure 4:26).

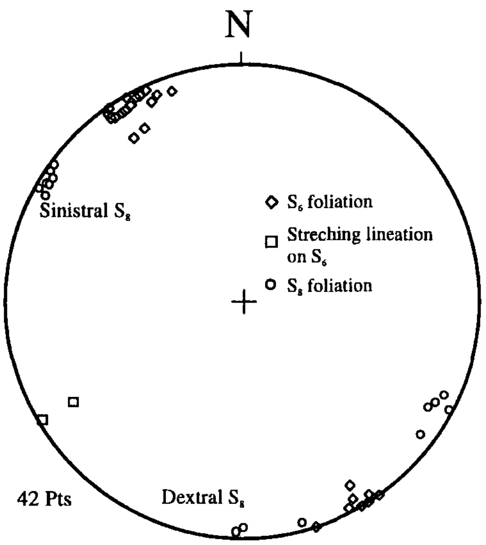


Figure 5:26
Stereogram of the deformation within the Dunlewy-Poisoned Glen area.

The width of the Thorr Granodiorite "strip" is 425 metres in this area, although it also contains rafts of pelite (of the Creeslough Formation) and to a lesser extent limestone and metadolerite. Sheets of biotitic and aplitic microgranitoids are also common and become thicker south-eastwards, although the typical widths are 0.5-2.0 metres. The contact of the Main Donegal Granite is observed on the SE side of the stream which flows from Sand Lough and Croloughan Lough at GR B 933187. Although the exact contact is not visible it can be constrained to within 25-30 metres based on the style of intrusion. Exposure to the immediate NW of the above grid reference shows autochthonous Thorr Granodiorite which itself is sheeted having intruded into pelite. In this area the presence of foliation-parallel mafic-rich microgranite dykes, similar to those seen in the Lackagh Bridge area, gives this country-rock a banded appearance (see figure 5:27a). In contrast, granite exposures situated 30 metres to the south show a much more chaotic picture with a multitude of pegmatitic intrusions of different sizes intruded into xenolithic rafts of Thorr Granodiorite of varying dimensions (see figure 5:27b). In addition to these pegmatites, the host Main Donegal Granite consists of fine to medium-grained granodiorites and a coarser grained, less

Figure 5:27a:- Autochthonous strip of Thorr Granodiorite along NW margin of the Main Donegal Granite which itself has sheeted into pelitic metasediments. Both of these rock types have been intruded by fine-grained MDG granodiorite sheets. Dunlewy-Poisoned Glen.

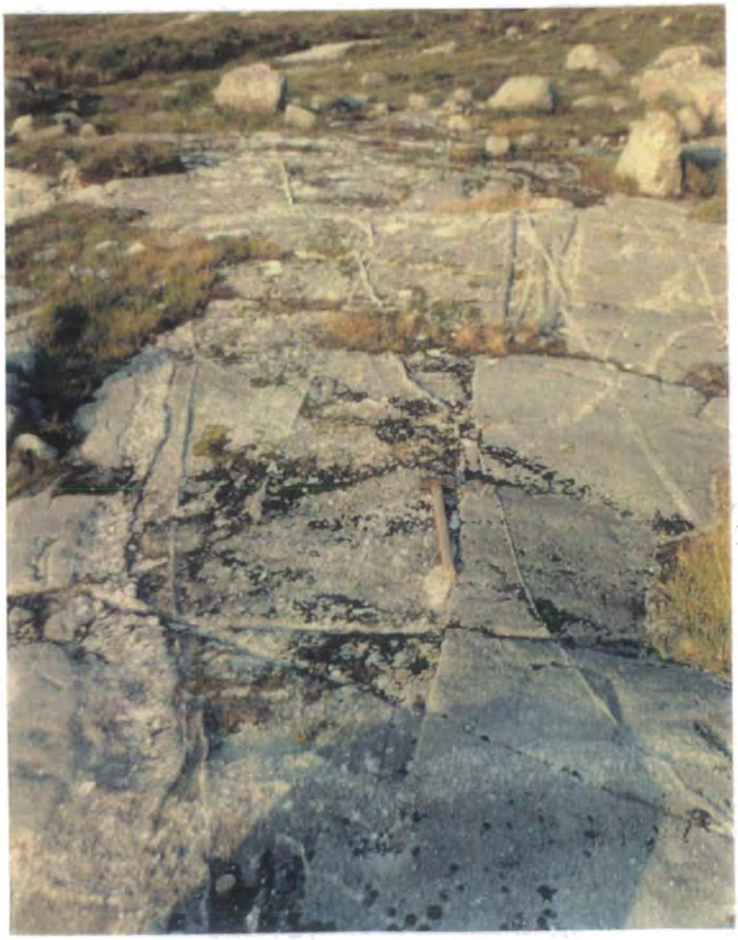


Figure 5:27b:- Composite xenoliths of Thorr Granodiorite, pegmatite and earlier MDG sheets broken up by later MDG facies. (30 metres away from fig5:27a). Dunlewy-Poisoned Glen.

biotitic monzogranite. The larger rafts of Thorr Granodiorite are often composite, consisting of pelite and occasional microgranitoid dykes which are commonly truncated against younger granitoids of the Main Donegal pluton. The smaller xenoliths tend to be smeared out and produce a streaky type of banding aligned parallel to the S_6 foliation. To the SE of this area, for a distance of 650 metres, these heterogeneous type granitoids are prevalent and show similar field relationships although the volume of pegmatitic material appears to increase.

A)

B)

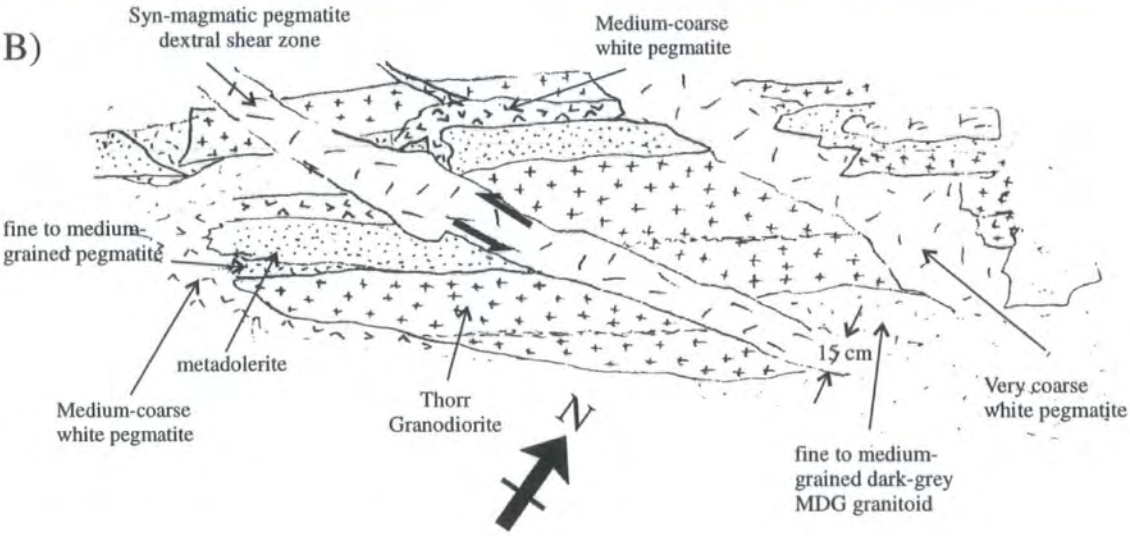


Figure 5:28:- (a & b) Heavily pegmatised marginal granitoids where a composite xenolith of Thorr Granodiorite, metadolerite, MDG facies and earlier pegmatite have been displaced by a syn-magmatic pegmatitic dextral shear zone.

At GR B 932179 this pegmatitic material is particularly abundant (see figure 5:28a & b) and intruded in to metadolerite, Thorr Granodiorite and MDG facies. In this figure there has been dextral displacement along the pegmatite vein probably during its intrusion. The characteristic E-W orientation, as seen in other dextral shears within the pluton, is consistent within an antireidel shear. The heterogeneous nature of this marginal granitoid facies disappears 150 metres to the SE of the above mentioned grid reference and passes in to homogeneous pink porphyritic monzogranites. Along the transect this relatively rapid transition between the marginal facies and the homogeneous pink porphyritic monzogranites is marked by the NE-SW trending gully at GR B 933178.

5:3:3:2 The homogenous pink porphyritic granites

Along the NW boundary of the porphyritic monzogranites the deformation is quite intense as indicated by well developed sinistral S-C' fabrics. Within these monzogranites there is very minor variation in regards to grain size and biotite content. Furthermore there are rare occurrences of localised banding which is aligned parallel to the foliation. At GR B 935173 there is a zone of Thorr Granodiorite xenoliths, up to 80 metres wide. These xenoliths are very angular and contain only very weak S-C' fabrics. To either side of this raft-train banding occurs within the pink porphyritic monzogranites in a zone up to 30 metres wide. Away from this small raft-train the banding dies out in to homogeneous porphyritic monzogranite.

The monzogranites between this small raft-zone and Lough Maumbeg (GR B 939161 are extremely homogeneous. The intensity of deformation within these monzogranites is considerably less than further to NW with only the S_6 foliation visible as indicated by aligned quartz on weathered surfaces. On the NW shore of Lough Maumbeg pelite rafts are encountered, accompanied with the appearance of banding in the porphyritic monzogranites. The width of this banding is approximately 125-150 metres and intensifies in a south-easterly direction. The SE boundary of this banding is the Derryveagh 2 raft-zone which is predominantly comprised of TG xenoliths. On the SE side of this raft-zone the granitoids are highly heterogeneous and are typical of the granitoids seen at the NW end of the Sruhanavarnis valley (as described in chapter 4). The well developed banding that is developed on the NW side of this raft-zone is not present with the granitoids along the SE side.

From studying the Derryveagh raft-zone in the Barnes Gap, Kingarrow and in the SE end of the Poisoned Glen it is apparent that this raft-zone has behaved as some form of barrier as the morphology of the granitoids and the style of intrusion on either side are very different. To the SE the granitoids are very heterogeneous, ranging from tonalites through to porphyritic monzogranites, whilst to the NW there are very uniform homogeneous pink porphyritic monzogranites. In the gully which joins the NW end of the Sruhanavarnis to the Poisoned Glen a vertical section through the Derryveagh 2 raft-zone is well exposed. In figure 5:29 it is clear that the angular Thorr Granodiorite xenoliths are immersed in a Main Donegal Granite host and separated from each other both vertically and horizontally.

Figure 5:29
A vertical section through the Derryveagh 2 raft-zone, showing its the angular xenolithic nature in the vertical plane. Gully between NW Sruhanavarnis and the "head" of the Poisoned Glen.

5:3:4 Brockagh-Meenderryherk

In this area the aim was to see if the pink porphyritic monzogranites are present to the immediate SE of the marginal facies as seen in other areas along the NW margin of the pluton. Figure 5:30 shows the distribution of the main granitoid types in this area.

At Brockagh the pink porphyritic monzogranite is encountered at GR B 847104 along the old track which runs SSE towards the Doocharry area. Again the NW boundary of this monzogranite with marginal facies is quite abrupt. The field relationships within the marginal facies are very similar to those described in the mouth of the Poisoned Glen, at the immediate contact, although at Brockagh rafts of pelite and metadolerite are more common.

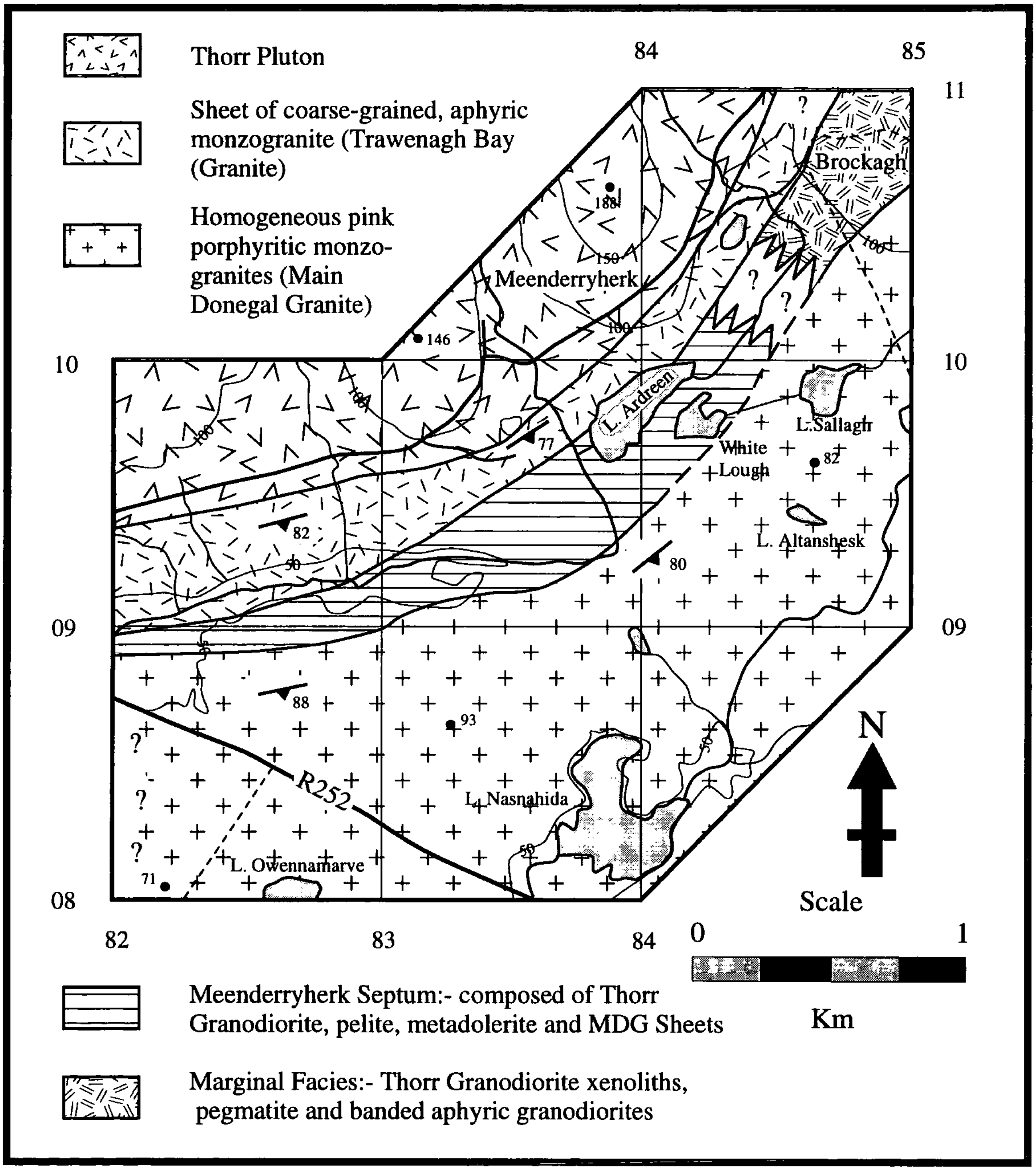


Figure 5:30:- The granitoids of the Brockagh-Meenderryherk area

Further to the SW, near the small hamlet of Meenderryherk, the transitional contact of the Main Donegal Granite and the Trawenagh Bay Granite occurs with both of these "Granites" being in sharp contact with the Thorr Pluton to the north. At Meenderryherk the homogeneous porphyritic monzogranites are encountered at GR 837092 close to the bend in a small river. The rocks to the immediate NW of this relatively sharp transition are composed of Thorr Granodiorite, pelite, metadolerite and small sheets of Main Donegal Granite although no xenoliths of Thorr Granodiorite or older MDG facies typical of the marginal facies were observed. Joint work by the present author and W.S. Pitcher has revealed that this mass of inclusion-rich, MDG sheeted Thorr Granodiorite is essentially a septum: the Meenderryherk Septum, and forms the NW boundary to the pink porphyritic monzogranites. The extent and significance of this septa will be addressed further in chapter 6. In the intervening ground between Meenderryherk and Brockagh the septum appears to break up possibly due to increasing densities of MDG sheets towards this latter area.

The pink porphyritic monzogranites of this area were compared with hand specimens obtained from the Poisoned Glen, Kingarrow and Lackagh Bridge and are almost identical in appearance (i.e. texture and modal proportions). The spatial extent of these granites in the Brockagh-Meenderryherk area is uncertain due to the relatively low amount of exposure, although for almost 200 metres from the NW boundary no variation was encountered. In the Doocharry area the banded tonalites and monzogranites, which form a synform, progressively die out in a NE direction (across the strike of the foliation) into homogeneous pink porphyritic monzogranites which are highly reminiscent of the porphyritic monzogranite seen at Brockagh.

5:3:5 Summary of the NW region of the Main Donegal Granite

It has been shown that the majority of the NW region of the Main Donegal Pluton is composed of homogeneous pink porphyritic monzogranites which are typified by their general absence of internal contacts and xenoliths of either country-rock or earlier granitic phases. What is clear is that these monzogranites form the largest homogeneous mass within the pluton, spanning its entire length, a distance of 35 kilometres. The SE boundary of this homogeneous monzogranite is the Derryveagh 2 raft-zone which separates them from the more heterogeneous granitoids which comprise the central regions of the pluton. To the SW of Slieve Snaght the relationship between the porphyritic monzogranites and the heterogeneous central zones becomes slightly less clear due to the Derryveagh 2 raft-zone becoming a less prominent feature.

From the dominant Thorr Granodiorite component within the Derryveagh 2 raft-zone it is apparent that the pink porphyritic monzogranites may have originally intruded into the "prolongation" of the Thorr Pluton. Whether or not the pink porphyritic monzogranites wedged apart the Thorr prolongation or were accommodated by passive mechanisms of emplacement is not clear from the study of the pluton alone. One question which does arise is the significance of the marginal granitoid facies seen along the NW, i.e. is it older or younger than the porphyritic monzogranites? Apart from the greater volumes of pegmatite in these marginal areas the biotite granitoids are very similar in appearance to the tonalites and granodiorites which comprise the central zones of the pluton. It was noted that in the Poisoned Glen section the marginal facies was very wide and pegmatitic material very abundant. To the SE of this, where the pink porphyritic monzogranites were encountered, the transition was sudden with almost no pegmatitic material present. This therefore might suggest that the pink porphyritic monzogranites are younger than the marginal facies.

5:4 The homogeneous zones as a whole

The lack of internal complexity within the granitoids of the Binaniller apophysis, Crockmore apophysis and the porphyritic monzogranites in the NW region of the pluton is in stark contrast to the central regions described in Chapter 4. The present author believes these homogeneous zones are portions of the pluton which have been constructed in a relatively rapid manner. The lack of internal contacts within these zones attests to this hypothesis. It is also believed that these zones are essentially younger than some of the granitoids which outcrop within the central areas of the pluton. As was mentioned in chapter 4 contacts between the porphyritic monzogranites were also very subtle and highly transitional in nature; a feature seen at Crobane Hill, in the Sruhanavarnis Valley and the Glendowan Mountains. It is therefore the authors belief that these porphyritic and aphyric monzogranites of the central regions may have been penecontemperaneous with the porphyritic monzogranites which comprise the homogeneous zones to the NW and SE, especially the pink porphyritic monzogranites and the Binaniller apophysis which are texturally, and to some degree compostionally, very similar. The relationship of the Crockmore apophysis which is more equigranular in texture, to the above mentioned monzogranites is less certain although it is believed that this granitoid is younger than the Binaniller apophysis and, therefore by inference, the porphyritic monzogranites. This may suggest that the Crockmore apophysis may be one of the youngest granitoid facies which comprises the Main Donegal Pluton.

Chapter 6

The relationship of the Main Donegal Granite to adjacent plutons within the Donegal Batholith

6:1 Introduction

The aim of this chapter is to address the relationship between the Main Donegal Pluton and the other plutons of the Donegal Batholith particularly with respects to intrusive-structural and temporal relationships. The common occurrence of fragments or xenoliths of older plutonic material within the outcrop of the Main Donegal Granite suggests that older plutons outcropped to varying extent within this area prior to its emplacement. Xenoliths of the Thorr Pluton are by far the most abundant of the older plutons which implies that this pluton may have been significantly larger than the present extent of its continuous outcrop. Furthermore, the presence of tonalitic xenoliths within the Carbat Gap area at the south-western part of the pluton suggests the existence of either the Ardara pluton or the Carbane Gneiss within the present outcrop of the Main Donegal pluton. The distribution of these older xenoliths will form the basis for the first part of this chapter. The remainder of this chapter will address the problem concerning the age relationship between the weakly deformed Trawenagh Bay Granite and the more deformed Main Donegal Granite. This work is based on detailed mapping of the granitoids which are situated along the presumed transitional boundary of these two "plutons". Furthermore joint mapping of parts of the Trawenagh Bay Granite with W.S. Pitcher (unpublished), has allowed some degree of correlation of the granitoid types seen in the highly complex transitional zone. The purpose of studying these other members of the batholith is to investigate whether or not the Main Donegal Granite Shear Zone, (MDGSZ) was in operation during the intrusion of the earlier plutons and to what extent it controlled the emplacement of later plutons.

6:2 The relationship of the Thorr Pluton to the Main Donegal Granite

The Thorr pluton is the oldest pluton of the batholith based on cross-cutting relationships (Pitcher & Berger 1972) and on radiometric absolute dating (O'Connor *et al.* 1982). The Rosses, Trawenagh Bay and Main Donegal plutons have all intruded areas of the Thorr pluton to varying degrees, although the present study has mainly addressed the relationship of the latter of these two plutons to the Thorr pluton.

The former extent of the Thorr pluton in the present area of the Main Donegal Granite has been inferred from the distribution of Thorr xenoliths within it (see Map D). Further mapping in certain areas (Cock's Heath Hill and Gartan Lough-Croaghacormick) has also helped in understanding the former extent of this earlier pluton.

6:2:1 The distribution of Thorr xenoliths with the Main Donegal Granite

The distribution of Thorr Granodiorite (TG) xenoliths within the Main Donegal Pluton has been determined from using the map compiled by Pitcher & Read (1959). It is noted that the majority of these xenoliths lie within, or to the NW of the Derryveagh 2 raft-zone. From Glen to Slieve Snaght, a distance of 25 km this relatively narrow raft-zone is amazingly continuous. It is this authors belief that this consistent raft-zone may have represented the southern contact of the NE "prolongation" (with the northern part of the prolongation preserved as an autochthonous strip along the majority of the length of the NW margin of the Main Donegal Pluton) of the Thorr pluton prior to the emplacement of the Main Donegal Granite. To the SW of Slieve Snaght the Derryveagh 2 raft-zone becomes less prominent and discontinuous with raft-trains of Thorr Granodiorite xenoliths now outcropping to the SE, across strike from this prominent raft-zone. In the well exposed ground (~GR B 917132) 1.25 km to the south of Crockfadda the occurrence of Thorr Granodiorite xenoliths within the Glenveagh 3A raft-zone is quite noticeable with composite rafts of TG, pelite and metadolerite present. It was observed that within the Glenveagh 3A raft-zone, 1.5 km to the NE of this area (i.e. the Bingorms), no xenoliths belonging to the Thorr pluton were observed. Therefore it is believed that the former contact of the Thorr pluton transgressed across the country-rocks in this area, as illustrated in figure 6:1 with the prolongation becoming somewhat wider to the SW. The occurrence of TG xenoliths in the Glenleheen area implies the contact was also present in this area and possibly connected with the autochthonous isolated mass of Thorr pluton preserved in the Cleengort-Lettermacaward area. The presumed southern contact of the Thorr pluton is therefore based on the distribution of TG xenoliths within the Main Donegal Granite. Berger (1967 *unpublished thesis*) came to a similar conclusion some thirty years earlier.

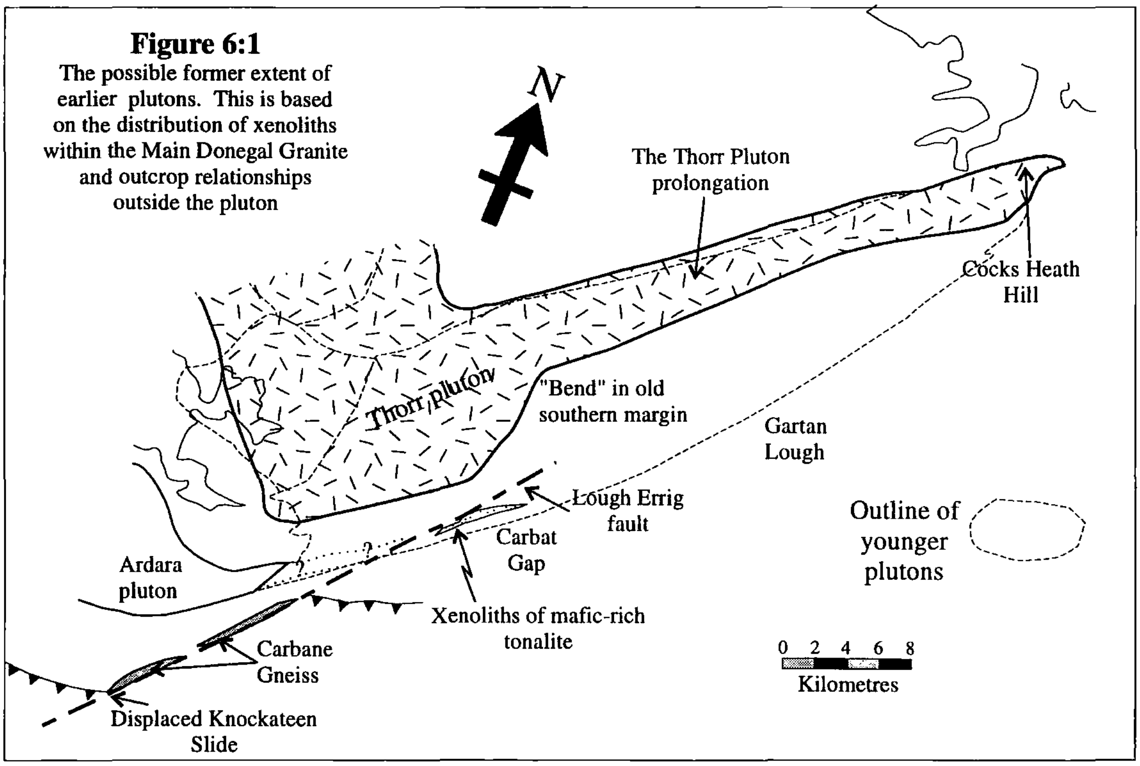


Figure 6:1
The possible former extent of earlier plutons. This is based on the distribution of xenoliths within the Main Donegal Granite and outcrop relationships outside the pluton

This southern contact of the Thorr pluton prolongation assumes that the majority of the raft-zones are dominantly in situ. One must consider also whether or not subsequent MDG granitoids have wedged apart the prolongation or have these later granitoids being accommodated by passive emplacement mechanisms. In chapters 4 & 5 it was stated that in the central areas of the pluton the granitoids are much more heterogeneous to the SE of the Derryveagh 2 raft-zone, whilst to the NW they are dominantly homogeneous. Therefore this raft-zone has behaved as some form of barrier during the emplacement of the Main Donegal Granite (see summary of section 6:2).

Along the NW margin of the Main Donegal Granite the autochthonous strip of the Thorr pluton is quite distinct. In a NE direction this "strip" thins until eventually at Cashel Mountain to the SE of Creeslough the Ards Quartzite forms the immediate envelope to the Main Donegal Granite, with Thorr Granodiorite now seen as xenoliths within the marginal facies as already described in Chapter 5. The present author looked for evidence of Thorr Granodiorite-meta-sedimentary contacts to the NE of this area in an attempt to ascertain the former north-easterly extent of the Thorr pluton "prolongation", i.e. does it extend further to the NE than the presently exposed Main Donegal Granite? Mapping in Cocks Heath Hill granite quarry and at the NE termination of the Main Donegal Granite itself has been undertaken to shed light on the north-easterly extent of the Thorr pluton "prolongation".

6:2:2 Cocks Heath Hill

Cocks Heath Hill is located 1.5 km to the NNE of Glen village and is situated quite near to the north-eastern termination of the Main Donegal Granite. Within the abandoned granite quarry at GR C 126327 (between the two roads) age relationships are well exposed. Within this quarry there are abundant Thorr Granodiorite (TG) xenoliths, semi-pelite rafts, fine grained microgranodiorite dykes and coarse grained pink porphyritic monzogranites. It is believed that this zone lies very close to the roof of the pluton as 0.65 km to the NE the Main Donegal Granite can be observed plunging beneath a flat-lying roof of Ards Quartzite. Within this quarry there is a fine example of syn-deformational banding as shown in figure 6:2 (a & b). From the exposure in the quarry itself it was difficult to tell whether or not the pelite forms rafts or is connected to the former roof of the pluton. What is clear is that this pelite is folded with the axial plane oriented 062°\87°SE whilst the plunge is 21°\058° and orientation which is parallel to the S_6 foliation, i.e. the fold was likely to have been produced by the shear zone (see figure 6:3). The banding in the adjacent granite may be produced by smeared out TG xenoliths in a host of pink porphyritic granite which also shows folding parallel to that in the semi-pelite (see figure 6:2a & b). The TG xenoliths are seen in contact with the semi-pelite implying the existence of Thorr material in the Cocks Heath Hill area prior to the emplacement of the Main Donegal Granite.

Within this small quarry, nearer to the entrance from the road, there is a well exposed sheet-like mass of tonalite-diorite which is dominantly intruded into semi-pelitic material (see figure 6:4a). The petrography of this rock is plagioclase (59%), biotite (20.4%), quartz (17.5%), K-feldspar (1.6%), muscovite (0.5%) and accessories (allanitic epidote, zircon, apatite and sphene) (1%). The high biotite content gives this rock a very dark-grey colour (fig. 6:4b) In comparison to the TG xenoliths within the quarry this rock is considerably more basic and is also more porphyritic due to the presence of plagioclase phenocrysts up to 10-15 mm in length. These laths tend to be euhedral and show well developed zoning and their faint alignment implies some degree of magmatic deformation (fig.6:4c). This rock has been subjected to considerable solid-state deformation with plagioclases commonly kinked and showing weak undulose extinction indicative of dislocation glide mechanisms. Quartz shows extensive evidence of dynamic recrystallisation with the majority of the grains quite weakly strained. Larger older grains are more flattened and subgrained and grade into small new grains, a feature indicative of SR recrystallisation. The presence of highly lobate quartz grains indicates that GBM recrystallisation has become more dominant during the later stages of deformation. The behaviour of quartz and feldspar implies quite high temperature deformation (400-450°C). This dyke is situated near to the

(A)

(B)

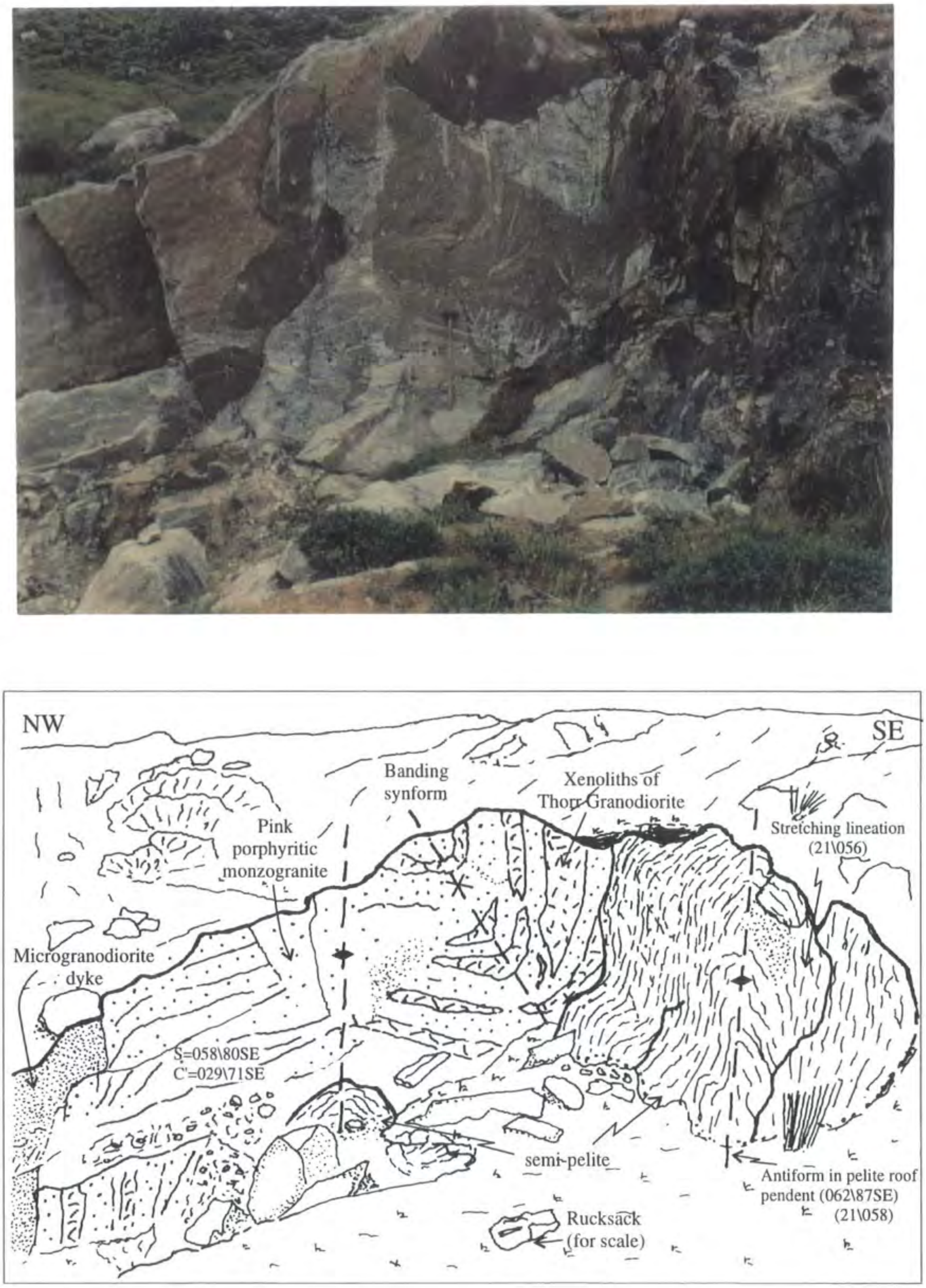


Figure 6:2:- Syn-deformational banding of Thorr Granodiorite xenoliths around folded semi-pelite within Cocks Heath Hill Quarry, (a) photograph, (b) field sketch (refer to text).

margin of the Main Donegal pluton and is intruded into pelites which are interpreted as roof septa extending down into pluton. This relationship implies that the diorite-tonalite is older than the MDG as corrobarated by the fact that granitic material resembling the MDG is intruded into it (see figure 6:4:a). Its relationship to the Thorr pluton is uncertain. To the SW of the quarry this sheet-like mass could be traced for up to 200 metres being observable in the summit region of Cocks Heath Hill. The significance of this intermediate sheet is uncertain, i.e. is it a more mafic component of the Thorr pluton or is it part of the relatively near Fanad Pluton? What is clear is that it is definitely not a granitoid typical of the Main Donegal Granite.

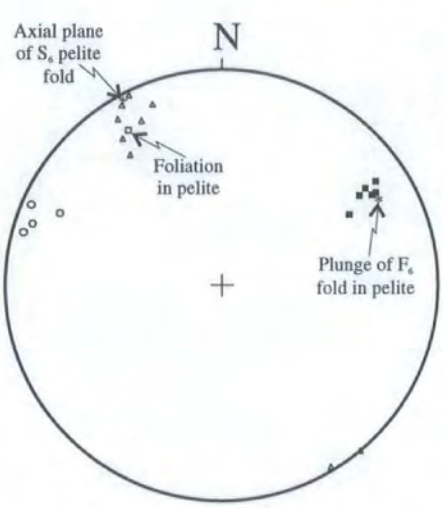


Figure 6:3
Structural data from Cocks Heath Hill granite quarry.

△ S_6 foliation

■ Stretching lineation on S_6

○ S_8 foliation

22 Pts

Figure 6:4a:- Large mafic-rich tonalite dyke at the entrance to Cocks Heath Hill Quarry.

Figure 6:4b:- polished hand specimen of the tonalite dyke seen in figure 6:4a.

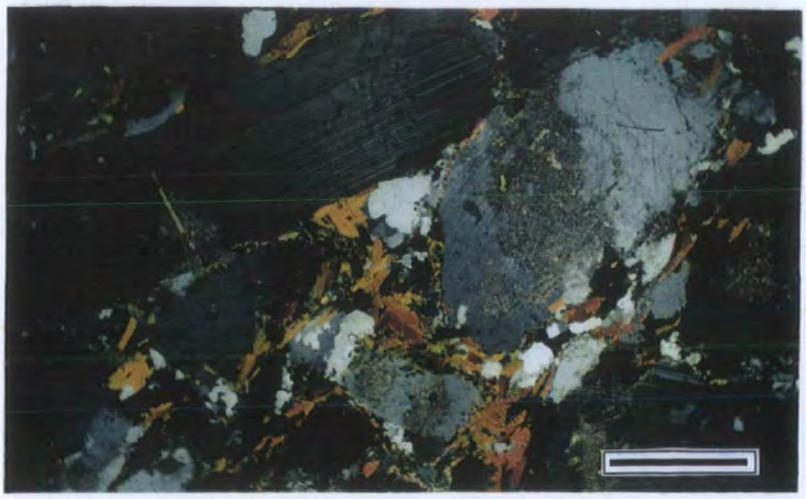


Figure 6:4c:- thin-section of mafic-rich tonalite in Cocks Heath Hill Quarry. It contains large plagioclase phenocrysts which show a moderate PFC alignment (scale bar is 2mm).

At the NE termination of the Main Donegal Granite itself the degree of exposure is very poor. Attempts were made to find autochthonous (i.e. in contact with the Dalradian envelope) granitic material belonging to the Thorr pluton to the NE of the termination of the Main Donegal Granite but nothing unequivocal was found. At GR C 133329 there are abundant boulders of Thorr Granodiorite and pegmatite, with some in situ material although no contact relationships between these and other granitoids or meta-sediments were observed. In this roof zone there appears to be localised stoping within the Ards Quartzite with randomly oriented xenoliths visible "floating" within a MDG granitoid host e.g. GR C 130331. Also small roof septa of quartzite are visible as the Main Donegal Granite fingers out in to individual sheets. In this area there are more shear zone folds which are axial planar to the S_6 foliation and plunge gently to the NE parallel to the stretching lineation. The granitoids of this area are banded aphyric granodiorites and pegmatites as seen in the marginal zones of the pluton (see chapter 5). In the quartzites to the NE of this termination no material of Thorr Pluton origin was encountered so it appears that this area corresponds to the NE termination of the Thorr Pluton "prolongation" and the Main Donegal Granite.

6:2:3 Gartan Lough

Berger (1967) claimed that there was no Thorr Pluton material to the south of the southern contact of the Thorr Pluton prolongation he drew across the Main Donegal Granite in his figure 88 (see figure 4:1 of this thesis for a similar position of this contact). The present author has found this not to be the case and has found xenoliths of Thorr Granodiorite (TG) in the Glendowan Mountains (Map I area; fig 4:34) and in the Gartan Lough area. In this latter area there is a considerable amount of TG material within the Main Donegal Granite, on the SE slopes of Croaghacormick and also along the margin of the pluton further to SE. The TG material seen in the Gartan Lough area is identical in appearance to TG xenoliths seen elsewhere in the pluton although here it is highly deformed (resembles an augen gneiss) due to its close proximity to the MDG margin. The TG along the sheeted margin of the Main Granite could be traced from GR B 034151 for 1.75 km to the NE to as far as Claggan Lough. Despite there being exposure of semi-pelite (Lower Falcarragh Pelite) and blackish-grey limestone (Falcarragh Limestone) in close vicinity to the exposures no unequivocal evidence to prove the exposures of Thorr were autochthonous in this area was found. Intruded in these Thorr Granodiorite xenoliths are considerable volumes of pegmatitic and aplitic sheets.

The nearest Thorr Granodiorite material to this area is in the Derryveagh 2 raft-zone, which at its closest, is encountered 6.5-7.0 km to the NW. In the intermediate ground no Thorr material was observed, despite the common occurrence of Dalradian sediments, i.e. the pelite rafts on Leahanmore-Croaghacormick and the Glenveagh 3A raft-zone to the immediate SE of Lough Veagh.

The significance of the Thorr Granodiorite in the Gartan Lough area is unclear due to it being so far from the main exposure of the pluton. Two options can account for this distribution:- i) the mass of TG in this area was originally intruded as an isolated mass away from the main part of the Thorr Pluton. ii) Originally this mass was connected to the main region of the pluton but has been tectonically disconnected from it possibly related to the emplacement of the Main Donegal Granite (see chap. 8).

6:2:4 Summary

The distribution of TG xenoliths within the Main Donegal Granite and the autochthonous strip along the NW margin is strong evidence for this pluton originally outcropping in this area before the intrusion of the MDG pluton. The continuity of the Thorr xenoliths within the Derryveagh 2 raft-zone implies that they are dominantly in situ and the former extent of the Thorr Pluton prolongation is as shown in figure 6:1. Of interest is the bend in the southern contact of this prolongation in the Crockfadda-Lough Barra area. There are two hypotheses which may account for this feature:-

- The bend is an original feature of the Thorr pluton where it transgressed across the Dalradian metasediments.
- The bend is only an apparent feature caused by the injection of more MDG granitic material in the area to the SW of this bend than in the area to the NE of it, i.e. the original Thorr Pluton has been wedged apart to a higher degree to the SW of the bend. Evidence which supports this model are the relationships seen on Crobane Hill. As was noted in chapter 4 the youngest and most voluminous granitoid phase in this area is the biotite-poor, equigranular CBH 3 monzogranite. In the Lough Slieve Snaght area it was noted that this monzogranite was in much lower abundance whilst in the Sruhanavarnis area it was not seen at all. On Crobane Hill the CBH 3 monzogranite is younger than the porphyritic monzogranites (CBH 2), a feature not seen in the other mapping areas where the porphyritic monzogranites are almost always the youngest. Thus there is evidence of a greater abundance of later material to the SW of this "bend".

Although the latter hypothesis may well have occurred, the author believes the "bend" is an original feature of the Thorr Pluton. The evidence which favours this is the presence of Thorr Granodiorite xenoliths in the Glenleheen area which lie along strike from the autochthonous mass of Thorr Granodiorite at Cleengort Hill and are therefore likely to be in situ. The origin of this bend within this old southern contact of the Thorr pluton may be a feature of the basement, i.e. fracture or lineament. This is further discussed in chapter 8.

The syn-kinematic evidence for the emplacement of the Thorr pluton into the Main Donegal Shear Zone is not particularly strong. The strongest evidence though is in the Poisoned Glen area where there are folds within semi-pelite which are axial planar to the S_6 foliation, i.e. the folds may be F_6, which have been truncated by a sheet of Thorr Granodiorite (see figure 6:5). Whether or not this fold was produced by shortening in the shear zone is uncertain, as it may have been of pre-Main Donegal Granite Shear Zone in age (i.e. pre D_6). If this fold was of D_6 age though it would demonstrate that this part of the Thorr pluton was syn-kinematic in relation to the shear zone.

Figure 6:5:- Possible F_6 fold truncated by a small sheet of Thorr Granodiorite, (NW autochthonous strip of Thorr, Dunlewy (GR B 930187).

The style of emplacement of the autochthonous Thorr strip in the Poisoned Glen area is very similar to that of the Main Donegal Granite, i.e. sheets intruded parallel to the S_6 foliation which implies there was some country-rock architecture controlling the emplacement of these two plutons although due to the high strains within these marginal areas there is also the possibility that sheets may have instead been rotated into the shear plane by subsequent deformation. Finding evidence for the syn-kinematic relationships of the Thorr pluton must involve study away from the zones of high strains that are associated with the deformation during the emplacement of the Main Donegal Granite as these lower strain areas would favour the preservation of earlier fabrics. Suitable areas may be within or at the margins of the Derryveagh 2 raft-zone or the south-western part of the Glenveagh 3A raft-zone.

Work by McErlean (1993) has shown that within the Thorr Pluton of the Crovehy Hills area magmatic state foliations are deflected in an anti-clockwise manner from their usual trend of NNW to a more WNW trend. This same author stated that deflection was in response to movement along the Main Donegal Granite Shear Zone, prior to the complete crystallisation of the Thorr Pluton. In a south-eastwards direction these magmatic foliations become overprinted by solid-state fabrics produced by progressive deformation related to movement along the MDGSZ. Therefore the southern part of the presently exposed Thorr Pluton can be regarded as syn-kinematic in respects to the shear zone into which the later Main Donegal Granite was emplaced.

Considering the overall shape of the Thorr pluton it appears to have been intruded in two main areas, controlled possibly by two different features. The main part of the presently exposed pluton is believed to lie within in a weakly developed NNW trending shear zone which Hutton & Alsop (1996) attributed to a deep crustal lineament. The Thorr "prolongation" on the other hand appears to have been controlled by a NE-SW trending feature, such as a possible precursor of a shear zone in which the Main Donegal Granite is now situated (see chapter 8).

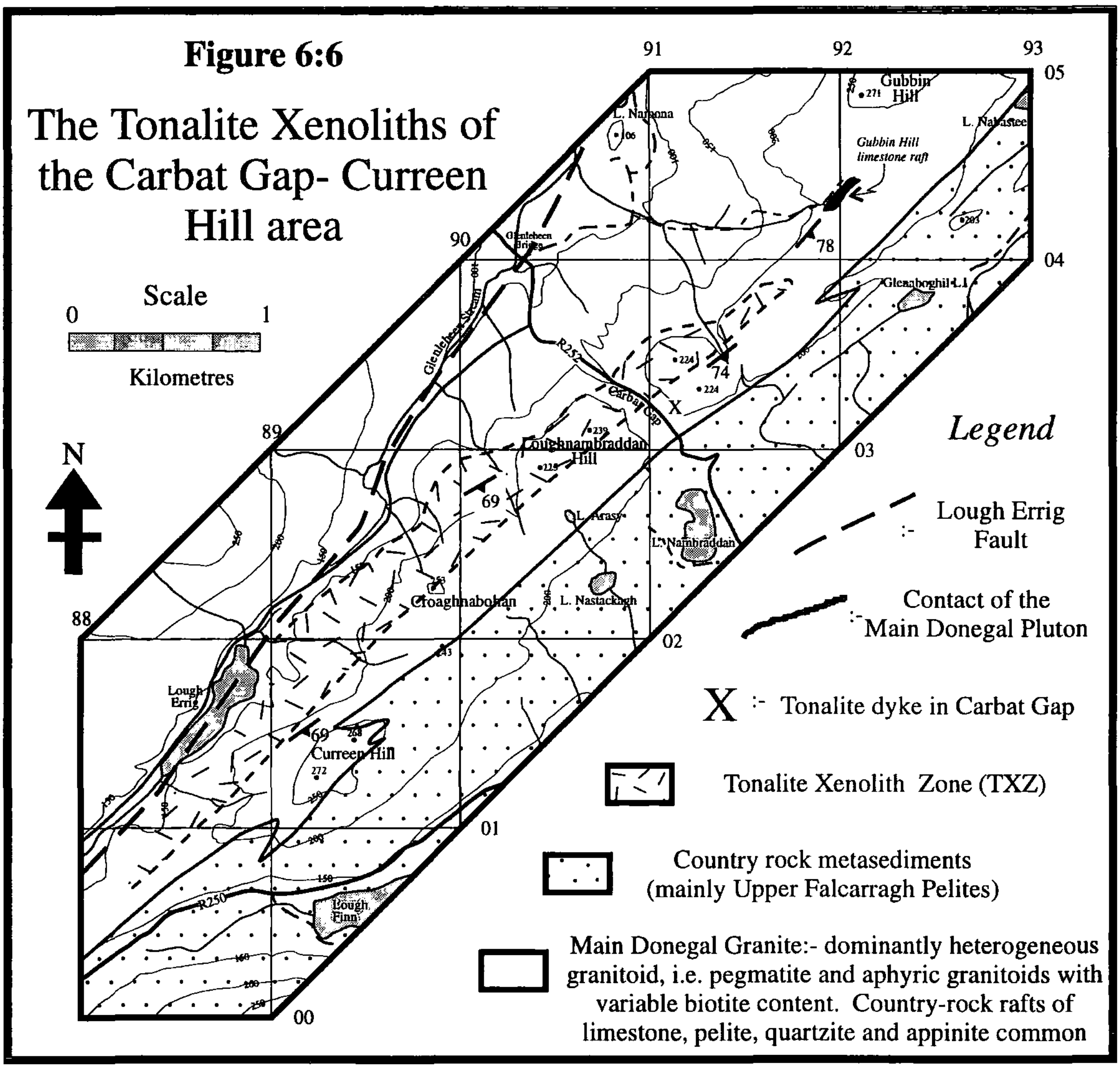
The following section discusses the occurrence of older tonalitic xenoliths within the more south-western part of the Main Donegal Granite.

6:3 The Tonalitic Xenoliths of the Carbat Gap

Introduction

In the vicinity of the SW margin of the pluton, between the small villages of Fintown and Glenleheen, there is a zone of highly angular and deformed xenoliths forming an area up to 250-450 metres wide and 4 km long (see figure 6:6). These xenoliths are most easily observed in a well exposed roadside section through the Carbat Gap. Cheeseman (1956), Pitcher & Read (1959) and Berger (1967) stated these xenoliths belonged to the outer member (G1) of the Ardara Pluton, whilst Pitcher (*pers comm.* 1996) mentioned the possibility of these xenoliths belonging to the Carbane Gneiss, an isolated tonalitic mass which outcrops to the south of the Main Donegal Pluton in the Glenties area. Berger (1967) stated that the forceful intrusion of the tonalite in the Carbat Gap area was responsible for causing a deflection of the country-rocks prior to the emplacement of the Main Donegal Granite, as now seen in the slightly oblique orientation of the Glenleheen 5 raft-zone preserved in the Main Granite. Hutton (1982) believed the more oblique orientation of this raft-zone was due to the lower strains within this part of the pluton, as compared with the higher strains in the more north-eastern portions, and hence the raft-zone had not been rotated into parallelism with the shear plane as much.

The following section will address the significance of the tonalitic xenoliths in this area based on a combination of field observations and thin section-analysis.



6:3:1 Field observations

The Tonalitic Xenolith Zone (TXZ) is situated close to the southern margin of the Main Donegal Pluton, only 220 metres away from it in the Carbat Gap area. The main contact of the pluton is visible 50-75 metres to the NW of the small roadside quarry at the south-eastern end of the Carbat Gap. The intervening granitoid between the contact and the TXZ is quite heterogeneous composed of pegmatite and banded aphyric granitoids which show fluctuating biotite contents. Common within these marginal type granitoids are rafts of pelite, metadolerite and quartzite which are well exposed on the roadside and on the southerly 224 metre summit (GR B 912033). Of particular interest within this marginal facies is the presence of a mafic rich tonalite which appears to be a sheet-like mass, up to 10 metre wide, intruded in to a pelite raft (GR B 911032). Its appearance is distinct from the tonalite which forms the TXZ and which outcrops to the immediate NW and which is also much more mafic than the adjacent Main Donegal Pluton granitoid facies (see polished hand specimens in figure 6:7 a & b for comparison).

Figure 6:7 a:- Polished hand specimen of a epidote-rich tonalite dyke which outcrops to the SE of the Tonalite Xenolith Zone (TXZ).

Figure 6:7b:- Polished hand specimen of the tonalite xenoliths which form the TXZ.

To the NE of the Carbat Gap this sheet-like mass could not be traced due to relatively poor exposure in this area. In the Carbat Gap area the TXZ is up to 250 metres in width occupying the higher ground, e.g. Loughnambraddan Hill and the northerly of the two 224 metre summits on the NE side of the road (GR B 911034). The host granitoids to the tonalites are dominantly coarse, pink pegmatites, and to a lesser extent fine-grained granodiorites and biotite-poor aphyric monzogranites. Within the zone of tonalites no large inclusions of country-rock material, i.e. pelite and quartzite, were observed. To the immediate NW of the TXZ there are prominent rafts of pelite and especially quartzite which trend parallel to the foliation within the pluton, i.e. ~057°. The host granitoid here is dominantly homogeneous aphyric granodiorite-tonalite (similar to the SRU 2 and GM 2 tonalites of Chapter 4) which contains occasional masses of appinitic material, (clearly visible along the Lough Errig-Glenleheen roadside e.g. GR B 898031) and also fine-grained inclusions of earlier Main Donegal Granite facies (granodiorite). These homogeneous granodiorites-tonalites are visible from the NW entrance to the Carbat Gap to Glenleheen Bridge. The granitoids to the NW of this point were not studied in detail, although previous work by Berger (1967) reveals that they become more heterogeneous with banded zones common particularly where in close vicinity to the various raft-trains which comprise the Glenleheen 5 raft-zone.

6:3:1:2 The internal nature of the TXZ

The xenolithic nature of this zone is particularly obvious in a near-vertical section through the Carbat Gap where inclusions of varying dimensions can be seen intruded by coarse, pink pegmatites (see figure 6:8). In the field the xenoliths are easily distinguished from the host granitoid due to the much higher colour index (~20%) and the presence of large plagioclase phenocrysts (up to 10 mm in length). Furthermore they commonly have a lightish algae growing on their surfaces. The contacts of the xenoliths with the host granitoids are generally very sharp and angular and imply the tonalite was highly competent during the intrusion of the Main Donegal Granite. In terms of exposure the xenoliths comprise up to 40-60% of the material within the TXZ and on the whole they are very homogeneous in appearance when compared with one another. However on the SE boundary of the TXZ though there is a narrow band of porphyritic diorite which in appearance is much more biotitic than the tonalite and also contains lower modal proportions of quartz (see polished hand specimen in figure 6:9) . Furthermore this diorite appears to be less deformed than the xenoliths typical of the TXZ, at least at the outcrop scale. On average the diorite was only between 5-10 metres wide and only found on the SE side of the tonalite zone. This diorite is relatively well exposed on Loughnambraddan Hill, to the NW of Lough

Arasy (GR B 905027) and on a small hill 300 metres to the west of Curreen Hill, in the Lough Errig area.

Figure 6:8:- The tonalite xenoliths in a vertical section through the Carbat Gap (xenoliths area covered with lightish coloured algae).

Figure 6:9:- polished hand specimen of the porphyritic diorite which forms a narrow strip up to 10-15 metres wide along the southern margin of the TXZ.

Subsequent deformation after the emplacement of MDG facies has caused localised welding of the tonalite xenolith-MDG facies contacts due to dynamic recrystallisation of quartz. Overall the xenoliths are intensely deformed and now resemble augen gneisses which on the outcrop scale appear considerably more deformed than the host Main Donegal Granite facies. The microstructural section will compare and contrast the strain preserved within the xenoliths when compared to the immediately adjacent host MDG facies.

It is apparent from the outcrop of the TXZ that prior to the emplacement of the Main Donegal Granite it possessed an approximate sheet-like geometry aligned in a NE-SW direction. Despite having been intruded by a plexus of granitoid facies related to the Main Donegal Granite this tonalitic zone has retained remarkably planar boundaries, a feature which the author believes is strong evidence for the TXZ being essentially autochthonous with only minimal displacements of the xenoliths during the intrusion of subsequent granitoid phases, i.e. pegmatite and aphyric monzogranites. If the above statement is correct then one might expect to see the preservation of tonalite-country-rock contacts within the rafts that now lie within the Main Donegal Pluton. It was this feature that the present author spent considerable time looking for on both the NW and SE boundaries of the TXZ. Along the highly planar SE boundary of the TXZ numerous rafts of pelite and quartzite within the marginal type granitoids could be seen to within 2-10 metres of the tonalite xenoliths although actual contacts are surprisingly rare probably due to later MDG granitoids exploiting such contacts. At GR B 911033 on the NW side of the peat bog which separates the two 224 metre summits this relationship is particularly clear with the wire fence situated on the TXZ boundary. At this same location, with the assistance of a spade, pelite and tonalite were seen in contact with one another. Along the NW boundary of the tonalite xenoliths a similar relationship exists where pelite and quartzite raft-trains are always seen in close vicinity, e.g. at GR B 906032. Towards Lough Errig, along the small road, there is a prominent cliff to the immediate SE of the road (GR B 895025) where original pelite-tonalite contacts are preserved for a distance of up to 100 metres and trending 058° (see figure 6:10).

In the SW the TXZ appears to terminate against the late-stage Lough Errig Fault which is situated along the Glenleheen Valley from Lough Errig itself to the Glenleheen area. Along the length of this fault all granitoid facies are extensively fractured and strongly discoloured with the original appearance of these granitoids almost impossible to determine. In this present study no xenoliths of tonalite resembling that of the Carbat Gap were observed to the NW of the Lough Errig Fault but one must emphasise that the degree of exposure in the vicinity of Lough Nabrack and Lough Namurleog is very poor. The NE end of the TXZ appears to be a natural termination with the tonalite tapering out into pelite and quartzite as marked by the general convergence of these country-rock rafts on either side of the TXZ. The splaying out of these rafts in a south-westward direction may confirm the belief of Berger (1967) that the tonalitic sheet forcefully intruded itself by wedging apart the Dalradian country-rocks.

Figure 6:10:- Photograph showing the preservation of tonalite-country rock (pelite) contacts (Glenleheen-Lough Errig road (GR B 895025).

6:3:2 Deformation in the Carbat Gap area

One of the striking features of the tonalite xenoliths is that they appear to be much more highly deformed than the host Main Donegal Granite. Commonly the xenoliths have well developed sinistral S-C' fabrics, particularly obvious due to the presence of large plagioclases, whilst the host granitoid has only a weakly to moderately developed S_6 foliation. At GR B 898026, 400 metres to the west of Loughnambraddan Hill, the tonalitic xenoliths are protomylonitic in character containing well developed δ-porphyroclasts. The host granitoid possesses a strong foliation but nothing as intense as the strain preserved in the tonalite xenoliths. If the deformation within the pluton was superimposed on to a cooling pluton as Pitcher & Berger (1972) had envisaged then the one would expect the xenoliths and the host granitoid to be equally deformed or possibly even the host to be more deformed than the xenoliths as it would have been more ductile as it crystallised and cooled. The more deformed nature of the xenoliths leads to the interpretation that they have been deformed for longer, i.e. in an earlier event. The latter statement seems more realistic and implies that the tonalite has been in the shear zone for longer and has therefore witnessed more D_6 deformation than the Main Donegal Granite which itself is deformed to a lesser extent by the same ongoing progressive S_6 deformation related to movement along the shear zone. There is no unequivocal evidence to imply that the tonalite of this area was syn-kinematic in relation to the Main Donegal Granite Shear Zone, (i.e. tonalite cross-cutting the S_6 foliation within the adjacent metasediments). However the tonalitic sheet-like mass at GR B 911032 within the marginal facies mentioned earlier does show syn-kinematic emplacement. To either side of the sheet are thin screens of pelite which have a well developed S_6 foliation (058\70SE) which has been truncated by the tonalite sheet. As the sheet cooled and the viscosity contrast diminished the sheet itself started to develop the S_6 foliation (053\67SE) (see figure 6:11). The relationship of this dyke to the TXZ or the Main Donegal Granite is uncertain. The more intermediate composition of this tonalite is not typical of any of the facies encountered within the Main Donegal Granite (apart from the large dioritic dyke seen in the Cocks Heath Hill Quarry). The significance of this sheet will be addressed in the summary below.

Figure 6:11:- The syn-kinematic nature of the small tonalite dyke to the SE of the TXZ (GR B 911032) where S_6 in the pelites has been truncated by the dyke, although within the dyke there is a well developed S_6 foliation which formed during its crystallisation i.e. both the pelite and tonalite being deformed by the same progressive deformation.

Despite the Main Donegal Granite appearing less deformed than the xenoliths within the TXZ, it has itself witnessed considerable deformation as indicated in the small quarry at the SE end of the Carbat Gap. This quarry lies within the sheeted country-rock, with the pluton contact situated 50-60 metres to the NW. Granitoid sheets of varying sizes have been boudinaged to varying degrees. It was observed that the sheets have undergone chocolate tablet boudinage, i.e. X=Y<Z. The orientation of X in the shear zone finite strain ellipsoid is approximately horizontal (plunges gently to the NE) and therefore would not produce such boudinage structures. It was this evidence which Hutton (1982) used to imply the Main Donegal Granite was essentially passively emplaced as up to 100 metres away from the contact the chocolate tablet boudinage is not seen within the deformed sheets.

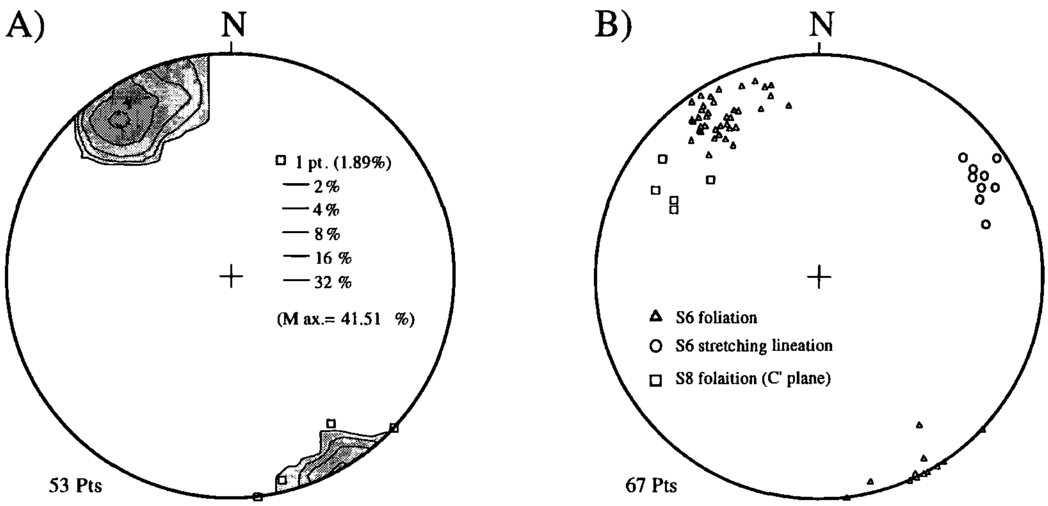


Figure 6:12:- Sterogram showing structural data from the Carbat Gap area. A) Contoured S_6 foliation plot. B) Composite plot of S_6, S_8 (C') foliation and S_6 stretching lineation.

Hutton (1982) claimed the during the intrusion of the Main Donegal Granite it developed an irrotational strain component of its own where X was vertical. The superimposition of the shear zone finite strain ellipsoid and the granite component led to bulk flattening within the pluton and the immediate aureole. If the granitoids of the pluton had been forcefully emplaced then the one would expect the sheets to be boudinaged for greater distances away from the contact. Hutton (1982) stated the component of strain within the granite was related to internal buoyancy and the confining pressure of the crystallising granite against the wall. As the pluton cooled to the ambient temperature of the country-rocks the granite strain component would expect to diminish and the pure flattening strains seen in the immediate aureole would be overprinted by continual movements along the acceptor shear zone. Within the quarry in the Carbat Gap the granitoid sheets have well developed stretching lineations and S-C' fabrics indicating sinistral shear which deformed the chocolate-tablet boudinaged sheets. This is an example of the overprinting nature of the shear zone during the later stages of emplacement of the Main Donegal Granite.

6:3:2:1 Thin Section

6:3:2:1a Tonalite xenoliths (XTZ)

In hand specimen these xenoliths resemble augen gneisses due to the presence of large feldspar porphyroclasts and generally appear considerably more deformed than the host granitoid. The behaviour of each mineral phase is as follows:-

1) Plagioclase:- forms large megacrysts, often zoned up to 8-10 mm in size and contain abundant fine-grained inclusions of biotite, epidote and to a lesser extent hornblende. Within the small size of the section these large plagioclase laths were aligned parallel to the foliation suggesting remnant PFC fabrics. Generally there is little evidence of any internal deformation within feldspar apart from slight kinking producing "sweeping" undulose extinction. No features such as brittle microfracturing were observed.

2) Microcline:- generally in low abundance within this rock and from textural relationships it is dominantly late stage and interstitial in distribution. All of this mineral phase is myrmekitised to varying degrees.

Generally there is little evidence of grain size reduction within feldspars such as marginal cataclasis or recrystallisation. Furthermore the feldspars are essentially quite fresh with only limited sericitisation.

3) Quartz:- forms a large part of the groundmass and is generally less than 2mm in size. There is abundant evidence of deformation related grain size reduction and dynamic recrystallisation. Generally grains are equant and show only weak undulose extinction and boundaries with other quartz grains are commonly planar, indicative of some degree of secondary recrystallisation. There are larger more deformed grains which show strong flattening to produce the S_6 foliation. It is these larger grains which tend to be highly strained and have well developed subgrains. Overall the dominant recrystallisation mechanism is grain boundary migration as indicated by the highly irregular, often lobate boundaries. The overall behaviour of quartz implies relatively high-temperature deformation (see figure 6:13).

4) Biotite:- the main mafic phase within this granitoid comprises as much as 20% of this rock and tends to have an olive-green colour and its strong alignment defines the foliation within this granitoid. Around the larger feldspar porphyroclasts biotite tends to be wrapped around causing minor deflection from the overall foliation trend.

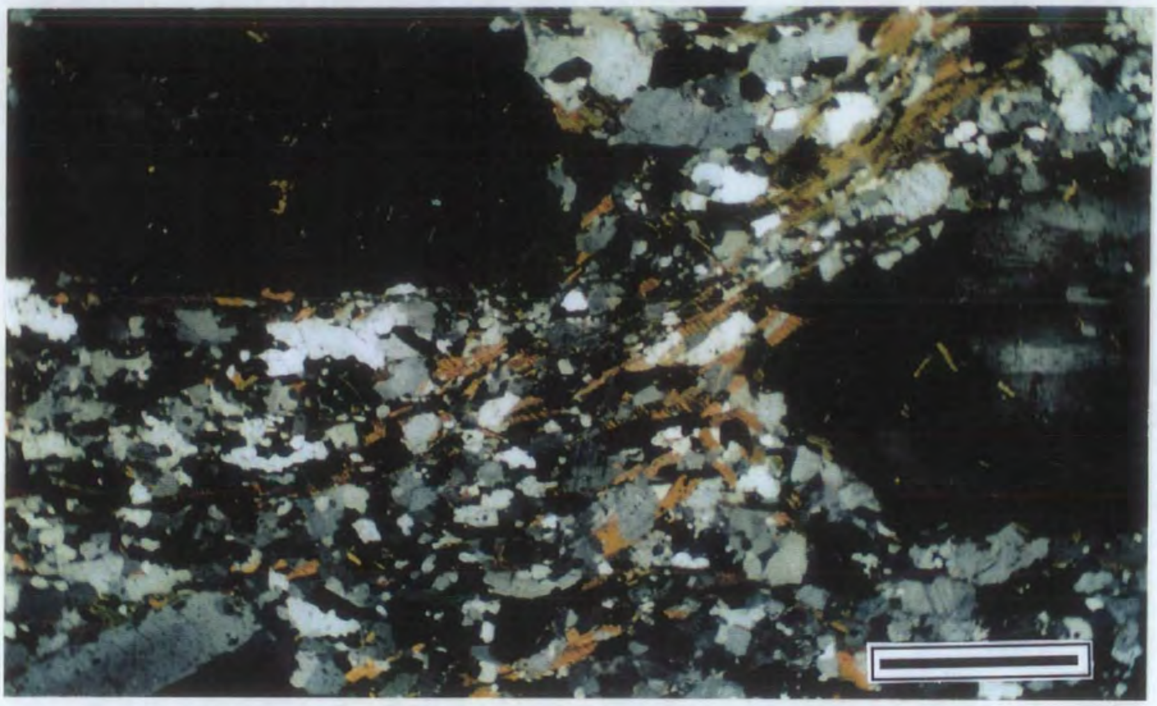


Figure 6:13:- Photomicrograph of tonalite xenolith. Large plagioclase phenocrysts show PFC alignment parallel to the S6 foliation (aligned biotites). Quartz occurs as small grains with lobate boundaries and appears strain-free indicating extensive dynamic recrystallisation (scale bar =2mm).

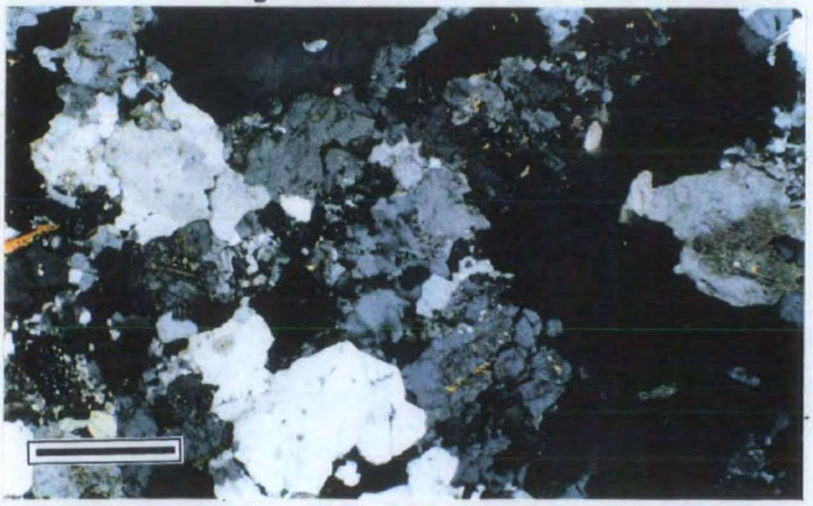


Figure 6:14:- Photomicrograph of the host MDG granitoid to the tonalite xenoliths. Large quartz crystals show strong undulose extinction although there is no evidence of any major dynamic recrystallisation (scale bar = 1mm).

5) Hornblende:- present to quite a low degree (<3% modal percentage) tending to be subhedral in shape and displays distinct pleochroism from lime-green to khaki-brown. Larger crystals have inclusions of quartz within them although no relict S*i* fabric is preserved.

Epidote and sphene form common accessories within these xenoliths, the former of which is seen in association with biotite or within plagioclase.

6:3:2:1b The host Granite to the tonalitic xenoliths

A number of samples of the host granitoids (aphyric monzogranite) was studied to compare the deformation between the tonalite xenoliths and the Main Donegal Granite (see figure 6:14).

1) Plagioclase is essentially undeformed and unaltered, showing no signs of internal deformation such as fracturing or sweeping undulose extinction. This phase has been extensively sericitised. There is no preferred orientation of plagioclase laths to indicate deformation in the magmatic state.

2) Microcline also shows no evidence of internal deformation although within all the samples it has been extensively myrmekitised (relatively low-temperature deformation).

3) Quartz shows the highest degree of deformation with all the grains showing very well developed undulose extinction with a minor degree of homogeneous flattening of this phase having occurred (see figure 6:14). Sub-grain development is relatively limited and evidence of dynamic recrystallisation is quite rare. Overall the quartz has been deformed by dislocation creep mechanisms and displays very little evidence of recovery.

4) Biotite:- despite the mica content being low this mineral phase shows no obvious preferred orientation in all the samples studied.

It is therefore apparent that the host Main Donegal Granite has been subjected to lower temperature and lower magnitude deformation than the tonalite xenoliths.

6:3:2:1c The Carbane Gneiss

This forms a series of fault bounded slivers in the Glenties area which extend for 6.5 km in a NE-SW direction and are usually less than 0.75 km wide. A sample of this material was collected to see if it bore any resemblance to xenoliths observed in the Carbat gap area (see figure 6:15 a & b). The overall petrography and deformation structures are as follows:-

1) Plagioclase has been deformed dominantly in a brittle fashion, with common microfracturing. Albite twins are usually bent or broken giving the grains a crudely

Figure 6:15a:- Polished hand specimen of Carbane Gneiss recording strong sinistral deformation (S-C' fabrics). Quartz is also intensely flattened forming the strong foliation.

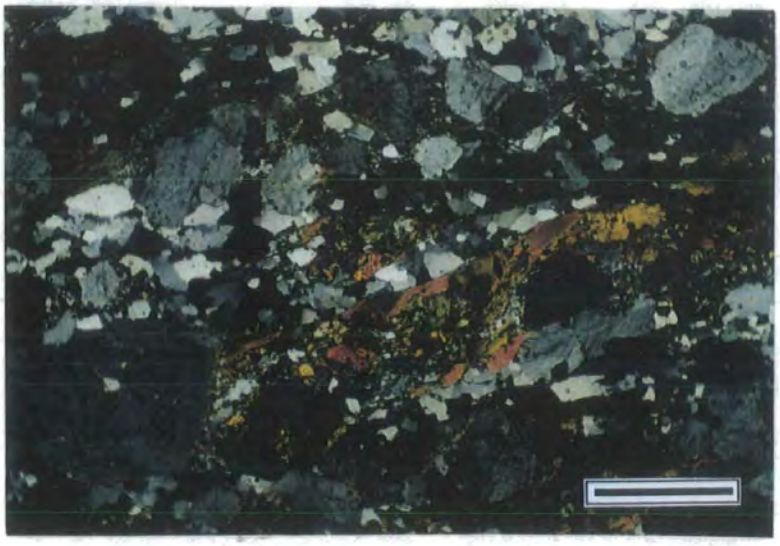


Figure 6:15b:- Photomicrograph of high-temperature solid-state deformation of Carbane Gneiss. Note the very fine-grained mantles of recrystallised feldspar around larger feldspar porphyroclasts. Quartz is extensively recrystallised showing "ribbon" textures consisting of fine-grained aggregates which appear relatively strain-free, i.e. no undulose extinction (Scale bar=1mm).

developed undulose extinction. Generally plagioclase crystals are up to 8mm in length and are more abundant when compared to the plagioclases within the Carbat Gap xenoliths. Overall these laths have a crude preferred orientation aligned parallel to the foliation suggesting some degree of magmatic-state deformation.

2) Microcline is generally uncommon in this rock although crystals up to 1 cm in diameter are observed containing inclusions of biotite and plagioclase. The majority of the microcline is interstitial in character.

Both feldspars around their margins have undergone limited degrees of recrystallisation (SR) with very fine-grained feldspar present. The uniform grain-size lobate boundaries argues against cataclasis of feldspar. This indicates that this gneissic tonalite has been subjected to high-temperature solid-state deformation.

3) Quartz tends to exist mainly as equant, weakly to moderately strained crystals and implies there has been extensive recrystallisation by both SR and GBM mechanisms. The grain size of quartz rarely exceeds 1mm in diameter. There are older grains which have better developed undulose extinction and show a higher degree of flattening and these tend to be larger than the recrystallised aggregates.

4) Biotite is very common in this tonalitic gneiss forming up to 20% of the mode and defines the strong foliation within this granitoid. It is common along both the S and C' planes and which show dominantly sinistral offsets.

Common accessories within this tonalite are epidote (up to 4%) and sphene. These are typically euhedral in form and have behaved in a predominantly brittle fashion with fractured sphene grains quite common.

6:3:2:1d The Ardara Monzodiorite

Samples of the G1 of the Ardara pluton were studied from the "stalk" of this pluton (samples obtained from S.J. Molyneux). These tend to be highly deformed probably due to movements along the Main Donegal Granite Shear Zone. Many similarities in petrography and texture were observed between this, the G1 granite, and the xenoliths at Carbat Gap (see figure 6:16). The G1 tonalite is porphyritic in appearance containing euhedral plagioclase laths up to 7-8 mm in size. Tiling features are present within these samples although whether these fabrics relate to the intrusion of the Ardara Pluton or the Main Donegal Granite within this area is uncertain due to the author not having studied this part of the Ardara Pluton. Microcline is relatively uncommon and formed late in the crystallisation history as indicated by its interstitial distribution. Biotite (16-18% modal proportion) is strongly aligned producing a strong foliation. Hornblende in places has been replaced to some degree by biotite and usually displays lime-green to khaki-brown pleochroism. Accessories include epidote and sphene.

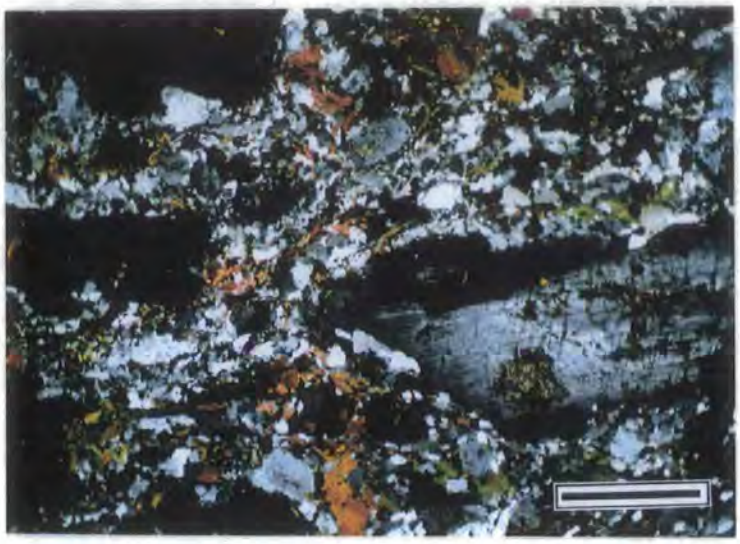


Figure 6:16:- Photomicrograph of a sample of G I taken from the "stalk" of the Ardara pluton. The presence of strongly aligned plagioclase phenocrysts and the extensively recrystallised quartz matrix is highly reminiscent of the tonalite xenoliths from the Carbat Gap (scale bar=1mm).

6:3:3 Summary of the Carbat Gap area

Overall the xenoliths most strongly resemble the G1 of the Ardara Pluton as Cheeseman (1956) had previously suggested. The Carbane Gneiss tends to contain more accessory minerals, notably epidote, and does not contain any hornblende (at least not in the sections studied). The xenoliths in the Carbat Gap are generally petrographically and texturally similar to G1 of Ardara, tending to contain large phenocrysts of plagioclase in a much finer grained host. In both areas hornblende is present plus similar percentages of accessory minerals although the xenoliths of the Carbat Gap are generally more quartz-rich when compared to the G1. The main feature is that the xenoliths are more highly deformed at Carbat Gap and also more deformed than the host Main Donegal Granite. An unusual porphyritic diorite which contains very little modal quartz occurs along the SE margin of the TXZ. This diorite was observed in the Carbat Gap (GR B 910034), 100 metres to the NW of Lough Arasy (GR B 905026) and on the small summit to the west of Curreen Hill (GR B 889014). To the immediate SW of these diorite rafts are meta-sedimentary rafts dominantly composed of pelite. This porphyritic diorite does not resemble any of the major facies comprising the Ardara Pluton (S.J. Molyneux *pers comm. 1997*) nor that of the Main Donegal Granite. In comparison to the Thorr Pluton this diorite appears to be more mafic and contains less quartz than typical samples of Thorr Granodiorite. The other alternative is that it may belong to the Carbane Gneiss as mafic contents are very similar, although the present author is not familiar enough with the entire outcrop of the Carbane Gneiss to test the validity of this statement.

Therefore it may appear that there is possibly granitoid material belonging to both the Ardara Pluton and the Carbane Gneiss outcropping in the Carbat gap - Curreen Hill area. What is clear that the xenoliths of Ardara were originally present in this area prior to the emplacement of the Main Donegal Granite as it was seen in contact with pelite along the Lough Errig road. The same may be true from the thin diorite strip along the south-eastern contact, although no direct contact with meta-sedimentary material was observed (despite it being constrained to within 5 metres). This argues against these xenoliths being transported from a different area during the emplacement of the Main Donegal Granite. Furthermore the general contacts of the xenolith zones are quite planar which suggests the existence of an autochthonous strip of tonalite in this area which may have been narrower than its present width after the pegmatite and medium-grained granitoidd of the Main Donegal Pluton are removed. Therefore the possible origins of this tonalite are as follows:-

A) Part of the Ardara Pluton was intruded as a sheet-like mass into the shear zone prior to the emplacement of the Main Donegal Granite. The "stalk" of the Ardara Pluton may be the remnant of the prolongation (see figure 6:1). Hutton (1982) claimed the stalk of the pluton was caused by shearing of the Ardara balloon by the cross-cutting shear zone. Molyneux & Hutton (1997 *in press*) have shown that during the late stages of crystallisation of the Ardara Pluton the shear zone was in operation. These same authors have not ruled out the fact that the Ardara Pluton may have ascended along part of the shear zone with the magma exploiting the shear zone to produce the stalk. Therefore the xenoliths in the Carbat Gap may be the remnants of an Ardara "prolongation" which existed in this area prior to the emplacement of the Main Donegal Granite.

B) The xenoliths of Ardara are generally deformed to a higher degree than the host Main Donegal Granite. In the xenoliths quartz is much more thoroughly recrystallised than in the Main Donegal Granite host either implying the xenoliths have been in the shear zone longer, i.e. have been deformed over a longer period, or the temperature of deformation was much higher.

C) If the deformation was superimposed on these granitoids then one would expect the xenoliths and the host to be equally deformed and possibly the host to be more deformed as it would be at a higher temperature as it fully crystallises and cools.

D) There may also be material resembling that of the Carbane Gneiss within the Carbat Gap area. The Carbane Gneiss is situated along a series of late stage faults, the most prominent of which is the Lough Errig Fault. To the south of the Main Donegal Granite this fault displaces the Knockateen Slide (see figure 6:1). Assuming that this fault has dominantly strike-slip displacements then restoration of the Knockateen Slide displaces the Carbane Gneiss close to the Lough Errig area. One important feature is that the contact of the Main Donegal Granite is not displaced by the Lough Errig Fault, but is deformed by it with brecciation common along the course of the Glenleheen Stream. This implies that if there is Carbane Gneiss material within the Carbat Gap then there must have been a large displacement along the Lough Errig fault and it must have been pre-Main Donegal Granite in age. The Carbane Gneiss has been subjected to high temperature solid-state deformation and the xenoliths of the Carbat Gap-Curreen Hill area are likewise highly deformed. Hutton (1982) claimed the deformation of the Carbane Gneiss was D_6 in age and represents the continuation of the Main Donegal Granite Shear Zone. The present author suggests the possibility of the existence of the Lough Errig Shear Zone (LESZ) where the majority of the displacement was achieved in a ductile fashion before the emplacement of the Main Donegal Granite. On the map of NW Donegal by Pitcher & Berger (1972) there is a deflection of the foliation in the MDG around the Lough Errig area. The sense of rotation of this foliation implies sinistral shear. The presence of rheologically weakened rock would favour reactivation by late stage faults during the Devonian. Therefore the Lough Errig shear zone may be possibly related to the Main Donegal Granite Shear Zone, i.e. it has a "R_1" shear geometry. Future work may involve looking for possible evidence of early ductile sinistral shear along the present line of the now dominantly brittle Lough Errig Fault between Glenties and Lough Errig itself, although the degree of exposure along the strike of this fault is quite poor.

6:4 The relationship between the Trawenagh Bay Granite and the Main Donegal Granite

6:4:1 Introduction

Previous research concerning the relative ages of these two plutons has already been reviewed in Chapter 3 of this thesis. The Trawenagh Bay Granite is dominantly intruded into the Thorr pluton, although the south-western portion of the pluton is intruded into Dalradian meta-sediments. As well as being younger than the Thorr Pluton the Trawenagh Bay Granite is also younger than the Rosses complex as sheets belonging to the latter are truncated by the former. The relationship of the Trawenagh Bay Granite to the Main Donegal Granite is less certain (as mentioned in chapter 3).

In this study the granitoids have been studied at two

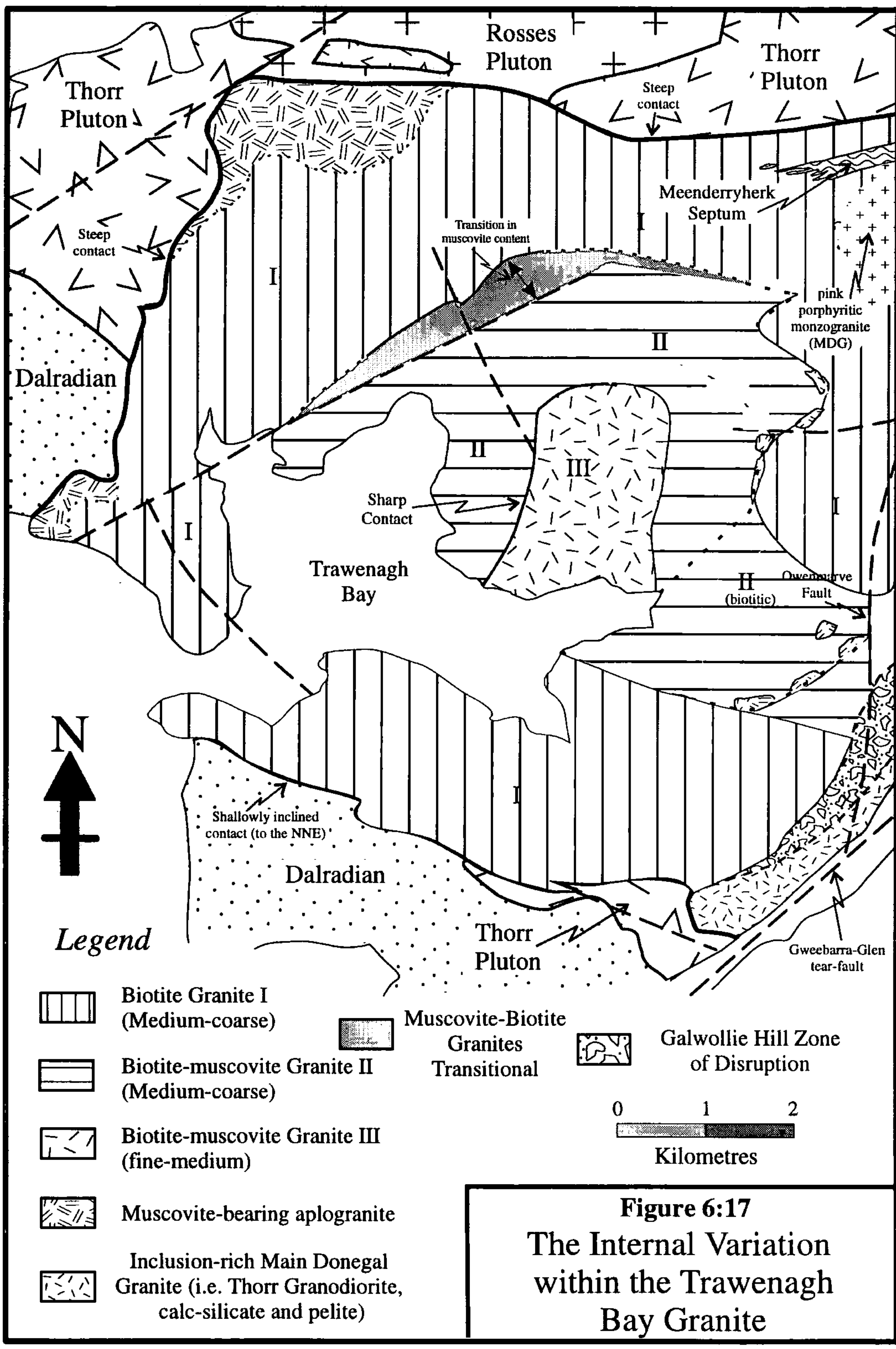


Figure 6:17
The Internal Variation within the Trawenagh Bay Granite

locations which are situated along their transitional boundary in an attempt to ascertain the age relationship between these two plutons: These two areas are Meenderryherk in the north and at Galwollie Hill to the south.

Recent re-mapping by Pitcher (unpublished) suggests that the Trawenagh Bay Granite is multi-pulse body consisting of up to three main units (see figure 6:17). The following description of this pluton is taken from Pitcher (1997 *pers. comm.*). The largest of the three units is a medium-coarse equigranular monzogranite-granodiorite which occupies the majority of the periphery of the Trawenagh Bay pluton. It is believed that this G I unit is a large sheet which dips towards the north. This is based on two lines of evidence; (i) the southern contact of the pluton in the Trusklieve area is gently inclined with the granite on top of the Dalradian meta-sediments. (ii) Towards the top of the sheet the G I become more muscovitic e.g. in the NE portion of the pluton there is a large area of muscovite-garnet aplogranite which is believed to form the upper part to the G I sheet. G II, a biotitic monzogranite (muscovitic to varying degrees) is smaller than G I and occupies the more central regions of the pluton. G II tends to become more muscovitic towards its outer contact which is generally quite steep and sharp. The contact between G I and G II is subtle as both facies are almost identical in appearance. Depending on weather conditions (preferably good sunlight) the contact can be located by observing biotite-muscovite monzogranites of G II juxtaposed against the biotitic G I as seen at GR B 807079. G III is the smallest pulse and occupies the central parts of the pluton and is typically a fine-grained monzogranite which also becomes more muscovitic to its margins. Pitcher reports that the outer contact of this pulse is intersheeted with G II. On the whole the three pulses of Trawenagh Bay area are internally quite homogeneous with xenolithic material quite rare. At one location (GR B 819023) near Ballynacarrick there are xenoliths of pelite, Thorr Granodiorite and MDG granodiorite within granitoid resembling G I.

This new re-mapping of the Trawenagh Bay Granite has been very useful as it allows a greater understanding of the granitoids which lie along this often complex transition between the Trawenagh Bay Granite and the Main Donegal Granite.

6:4:2 Meenderryherk

This area was studied jointly with W.S. Pitcher and the aim was to attempt to trace the Border 1 raft-zone (NW "marginal facies" of this thesis) from the Main Donegal Granite into the Trawenagh Bay Granite. Previous research by Pitcher & Read (1959) claimed that this raft-zone lost its strong NE -SW alignment and broke up as it entered the Trawenagh Bay Granite and it was this evidence that suggested the Trawenagh Bay Granite was younger than the Main Donegal Granite, or at least the deformation suffered by this pluton. New work has shown that this so-called raft-zone is actually a septum composed of Thorr Granodiorite, pelite, metadolerite and sheets of Main Donegal Granite. At Meenderryherk this septum is up to 500 metres wide (see figure 5:30). To the NE the septum, as mentioned in chapter 5, it becomes less obvious due to increasing amounts of pegmatitic material so that it starts to resemble the MDG marginal granitoid facies of the NW margin of the pluton which contains Thorr Granodiorite xenoliths. At Meenderryherk this mass of sheeted Thorr Granodiorite was observed to be more continuous than previously thought, i.e. a septum rather than a raft-zone traceable into the Trawenagh Bay Granite to as far west as the Lough Sallagh Bridge (GR B 816090). In this area the septum rapidly thins and fingers out into homogeneous GI granodiorite/monzogranite belonging to the Trawenagh Bay pluton. It was observed that the septum does change orientation when traced westwards and swings from NE -SW into parallelism with the northern boundary of the Trawenagh Bay Granite which trends E-W. To the immediate SW of the Meenderryherk Septum in the Brockagh and Meenderryherk area there are homogeneous pink porphyritic monzogranites which are seen along the entire length of the NW part of the Main Donegal Pluton. To the NW of the same septum there is a homogeneous white biotitic monzogranite which forms a sheet up to 350 metres wide at Lough Sallagh Bridge but progressively thins in an easterly direction. Accompanying this thinning is an overall decrease in grain size from 3-5 mm to 1-3 mm. This sheet of monzogranite was traced as far NE as Brockagh, where relationships become more complicated by the increasing abundance of marginal granitoid sheets and pegmatite (see figure 5:30). Along the side of the road from Meenatotan to Meenderryherk (GR B 825094) the contact between the Thorr pluton and the sheet of white monzogranite can be easily traced and where directly observable the contact is knife sharp. At the above mentioned locality marginal sheets are rare to absent. The present author and W.S. Pitcher concluded that this white monzogranite sheet, which becomes finer grained to the NE, in fact belongs to the Trawenagh Bay Granite. The decrease in grain-size is interpreted as a chilling effect of the Trawenagh Bay Granite intruding in to the cooler Main Donegal Granite. Therefore it appears that at least the northern part (if not all of the pluton) of the Trawenagh Bay Granite is younger than the Main Donegal Granite as this large sheet intrudes in to it.

At Meenderryherk the deformation within all rock types is intense. In the Thorr Granodiorite at GR B 837100 there are excellently developed S-C' fabrics indicating sinistral shear. The granitoids 400 metres to the SW of this location are much less strained with the C' (S_8) foliation starting to disappear, whilst the S_6 foliation swings in to a more east-west direction, parallel to the northern boundary of the Trawenagh Bay Granite. It must be emphasised that the sheet of Trawenagh Bay Granite is also intensely deformed as it enters the Main Donegal Granite Shear Zone.

6:4:2:1 Summary

In the Meenderryherk area it has been observed that the Border 1 raft-zone does not break-up as it enters the less deformed Trawenagh Bay Granite, instead in the Meenderryherk area it forms an autochthonous septum (the Meenderryherk Septum) composed of Thorr Granodiorite, meta-sediments and sheets of Main Donegal Granite (although these sheets become less abundant in a westward direction). In a NE direction towards Brockagh this septum becomes progressively more impregnated by granitoid sheets and pegmatite belonging to the Main Donegal Granite and it was probably the increased concentration of these sheets which ultimately caused it to break-up into xenoliths. It is clear that the Meenderryherk Septum lies across the shear zone boundary as deformation within the septum clearly decreases along strike in a westerly direction. To the immediate NW of this septum there is a sheet of homogeneous monzogranite which belongs to the Trawenagh Bay Granite and appears to have wedged apart the Thorr Pluton to produce the Meenderryherk Septum. As mentioned this septum contains MDG marginal sheets which decrease in abundance towards the west. It was observed that within the Trawenagh Bay Sheet there were no marginal sheets present which implies this large sheet is later than the MDG sheets. This therefore agrees with the findings of Pitcher & Read (1959) who observed truncation of sheets along the northern contact of the Trawenagh Bay Granite. The absence of granitoid sheets in the western part of the septum would imply that the Main Donegal Granite did not occupy this area prior to the emplacement of the Trawenagh Bay Granite, i.e. the zone of maximum deformation corresponds to the original western contact of the Main Donegal Granite.

6:4:3 Galwollie Hill

Galwollie Hill is situated to the north of the Gweebarra Estuary and is most easily accessed from the Lettermacaward-Doocharry road (see figure 6:18 for location). The exposure is excellent with this area lying along the general transition between the highly deformed Main Donegal Pluton and the more weakly deformed Trawenagh Bay Granite. Mapping has shown that the granitoids of Galwollie Hill itself lie within a zone of very high strain corresponding to the north-western boundary of the Main Donegal Granite Shear Zone. This deformation tends to decrease both to the NE and SW of the Galwollie Hill area.

On the SW side of Galwollie Hill (350 metres from the summit) detailed mapping on a scale of 1:250 was performed (see Map C of this thesis). The method of mapping was similar to that already described in the Crobane Hill and Sruhanavarnis mapping areas. In comparison to these two latter areas the continuity of exposure is much less, although it is sufficient to obtain the chronology of the granitoids present and to give an insight in to the overall style of emplacement.

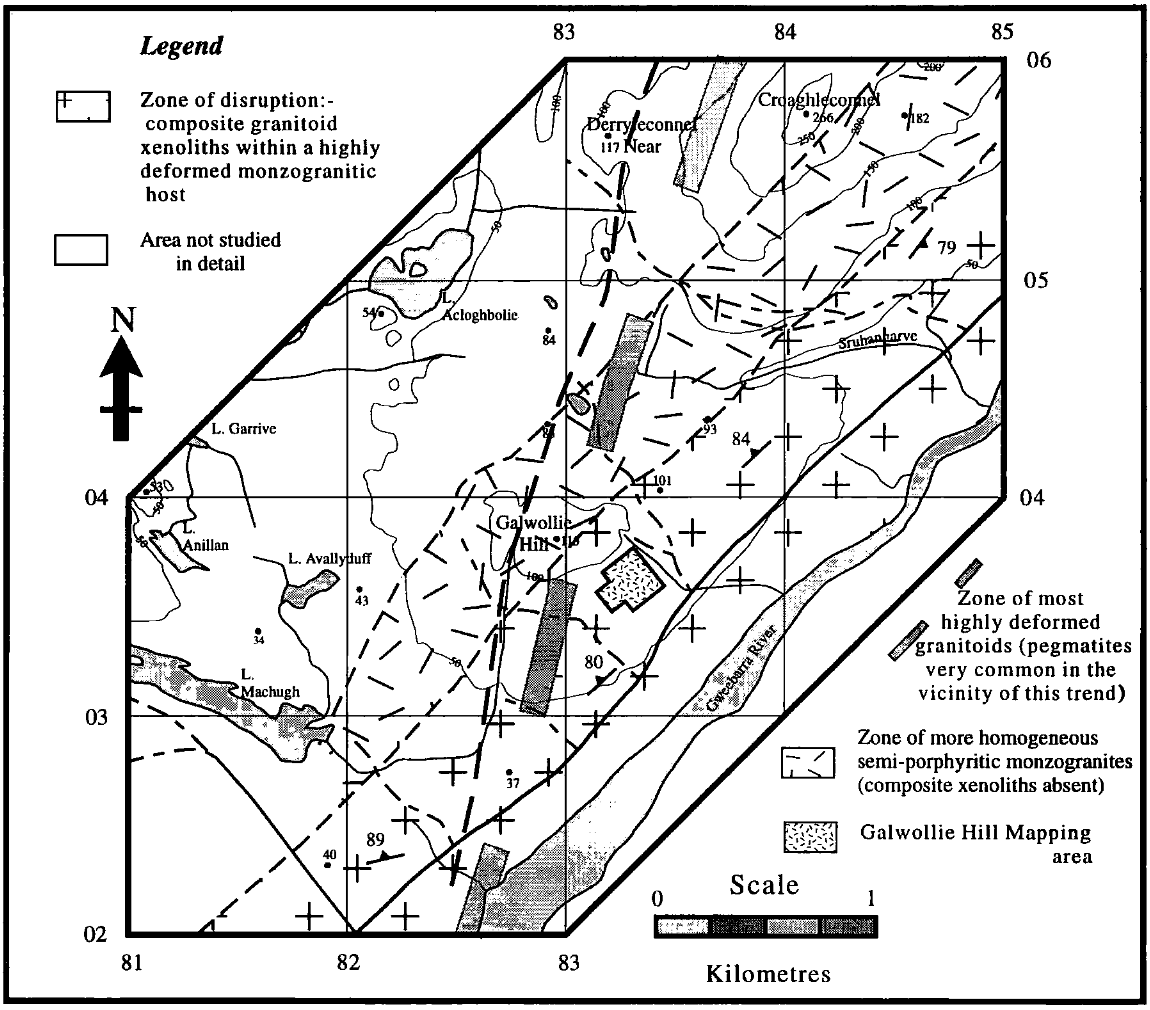


Figure 6:18:- The location of Galwollie Hill

On Galwollie Hill there are up to six or seven granitoid facies present (see Map C for distribution of these granitoids). The oldest facies is the Thorr Granodiorite which occurs as xenoliths in this area. The presence of autochthonous Thorr Granodiorite in the Crohy Hills, at Lettermacaward and on Cleengort Hill implies that the Thorr Pluton originally occupied this area prior to the emplacement of the Trawenagh Bay and Main Donegal Granites. The other granitoids of Galwollie Hill range in composition from tonalites through to quartz-rich monzogranites. Intruded in this area is a considerable suite of coarse pegmatitic material which Pitcher & Read (1959) interpreted as a roof partition, separating the two plutons, which dies out at depth. The aim of this section is to describe the granitoids which outcrop in Map C and discuss their overall relationships. In later sections the granitoids will be correlated with those seen in other parts of both the Trawenagh Bay Granite and the Main Donegal Granite. Finally the significance of this area will be addressed.

6:4:3:2:- The Granitoids of Galwollie Hill

The granites of this area will be addressed in order of age and will be prefixed by GWH (after Galwollie Hill). The oldest granitoid in this area is the Thorr Granodiorite and is quite distinct from the other granitoids. In general the large euhedral phenocrysts of plagioclase, the high biotite content and the distinct greenish-grey appearance on weathered surfaces allows it to be distinguished from the lighter grey, less mafic facies. Photographs of polished hand specimens are within Appendix A of this thesis. The modal compostions are plotted on a QAP plot as shown in figure 6:19).

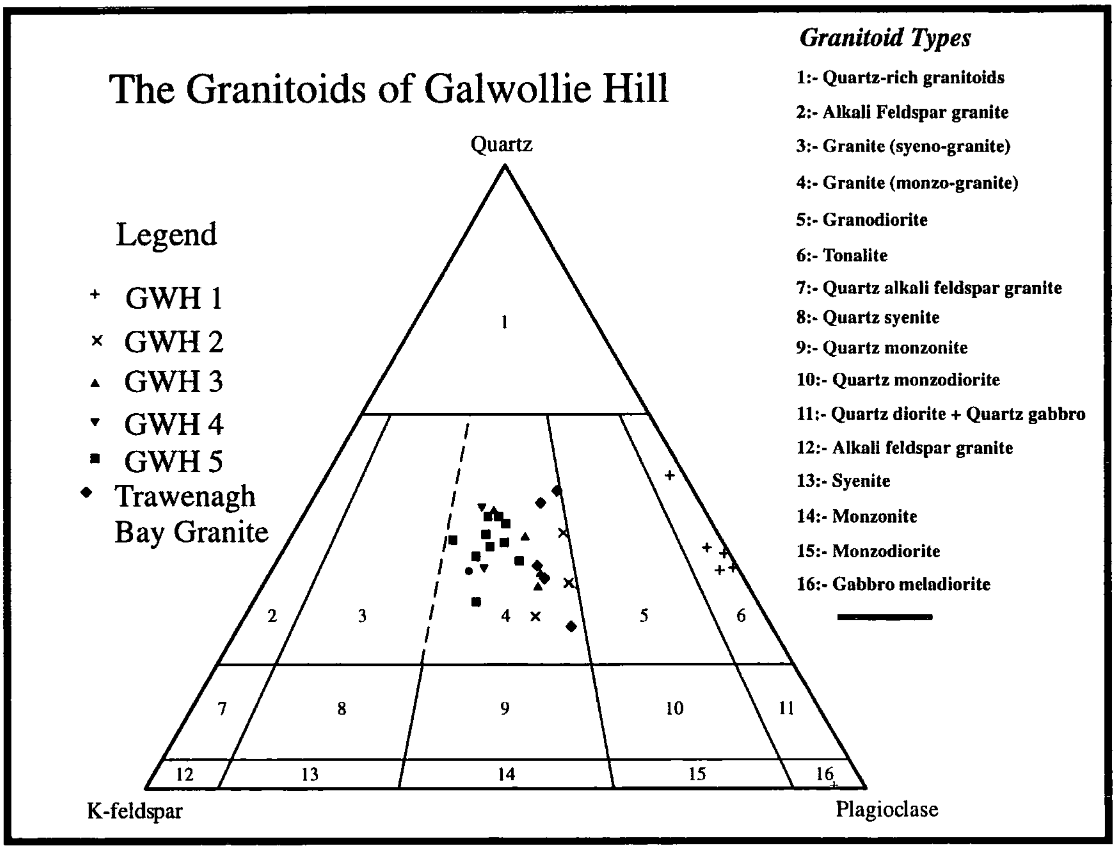


Figure 6:19:- QAP plot for the granitoids of the Galwollie Hill area.

GWH 1:- fine to medium-grained tonalite with the grain size generally less than 3 mm, although there are euhedral zoned laths of plagioclase up to 5 mm in size. The main components of this tonalite are white plagioclase (54%), quartz (31.5%), khaki-brown to yellow biotite (12.5%) with accessory muscovite (1%), epidote (0.2%) and apatite (0.1%). K-feldspar, in the form of microcline is less than 5% in modal proportions and is quite often absent. The high mafic content of this tonalite (~13%) gives it a dark-grey appearance on weathered surfaces. Apart from microgranitoids of varying ages the GWH 1 tonalite is the finest grained facies in this area. No variation within this tonalite was encountered within the mapping area.

GWH 2:- a porphyritic monzogranite which has a grain size of 2-4 mm although the megacrysts of K-feldspar (26%) can be as large as 6-7 mm. The overall colour of fresh surfaces of this rock is pinkish-white due to a very subtle pink discolouration of the K-feldspar megacrysts. Other components are white plagioclase (38%), quartz (27.5%), olive-green biotite (7.5%) and to a lesser extent muscovite (1%). Weathered surfaces are darkish grey, although the "chalky" white weathering nature of K-feldspar megacrysts allows it to be distinguished from the GWH 1 tonalite. Overall this monzogranite is very uniform in appearance.

GWH 3:- this is another porphyritic monzogranite but in comparison to the GWH 2 monzogranite the groundmass is generally much coarser (3-5 mm) and this monzogranite contains larger, pinkish-red K-feldspar megacrysts (26%). The texture of the GWH 2 and GWH 3 monzogranites are clearly different and are not differently weathered varieties of the same facies. This monzogranite is composed of red microcline, white plagioclase (33%), quartz (32.5%), khaki-brown to yellow biotite (6%), muscovite (2%) (mainly secondary alteration of plagioclase) and accessory epidote and minor apatite (0.25%). Within GWH 3 the biotite content does fluctuate within the area of Map C (between 7.5-10%). On weathered surfaces this monzogranite has a distinct light-grey colour and porphyritic appearance which allows it to be differentiated from the slightly darker and finer grained GWH 2.

GWH 4:- a coarse-grained biotite-muscovite, weakly porphyritic monzogranite. The porphyritic nature of GWH 4 is best appreciated in thin section where large megacrystic crystals of poikilotopic orthoclase and microcline (32.5%) are visible up to 8 mm in length. Other components are quartz (34.5%), plagioclase (26%) biotite (3.5%), muscovite with accessories relatively uncommon (<0.5%). The overall grain size of this monzogranite is 3-6 mm. The most striking feature of GWH 4 is the occurrence of muscovite flakes (up to 1.5 mm in diameter). This muscovite appears to be of primary nature and comprises up to 4% of the mode. Fresh surfaces tend to have a yellowish-pink colour to them, although the more highly deformed GWH 4 monzogranites develop a pinkish colour. On weathered surfaces GWH 4 has a light-grey colour. It can be distinguished from the GWH 3 monzogranite by its less porphyritic appearance and, if there is good sunlight, the muscovites can clearly be seen.

GWH 5:- this is the most heavily deformed and most common granitoid in the Galwollie Hill mapping area and tends to have a deep reddish-pink colour on fresh surfaces. The strong red colour is deformation related. On weathered surfaces quartz shows very strong alignment with C' shear band cleavage (S_8) ubiquitously developed. This quartz-rich monzogranite has a non-porphyritic appearance in hand specimen and contains grey quartz (36%), pinkish-red K-feldspar (32%), white-pink plagioclase (25%) olive-green biotite (4%) and muscovite (4%). Accessory minerals are relatively uncommon within this phase (<0.25%). The grain-size is variable with considerable deformation related grain size reduction. Weathered surfaces have a light-grey colour due to the lower biotite content and the presence of large muscovite flakes allows this monzogranite to be easily identified from the other granitoids. Comparison of hand specimens and thin-section analysis has led the author to suggest that the GWH 5 monzogranite may be a more highly deformed variety of the GWH 4 monzogranite as the chemistry of these granitoids is very similar. In Map C the GWH 4 and GWH 5 phases were differentiated by their appearance on fresh surfaces of hand specimens.

Pegmatite

Within the Galwollie Hill area there is abundant pegmatitic material. This zone of pegmatite extends from Brockagh in the north, through Croaghleconnel, the present mapping area and through Derkbeg-Straboy along the south-western margin of the Main Donegal Granite. This pegmatite is composed of very coarse grained microcline, quartz, muscovite, occasional biotite or garnet. The size of feldspars can be up to 5cm in diameter. It is apparent that there are two phases of pegmatite in this area, the earlier phase of which is the most voluminous and forms large masses that are mostly, but not always, confined to the Thorr Granodiorite, GWH 1, GWH 2 and GWH 3 granitoids. A lesser, more regular phase of pegmatitic intrusions is observed intruding through all of the main granitoid phases on Galwollie Hill.

6:4:3:3 Age relationships between the Galwollie Hill granitoids

The granitoids of Galwollie Hill are uniquely typified, within the MDG by angular composite xenoliths within a host of progressively younger monzogranites. Cross-cutting relationships are generally excellently preserved in this area allowing the relative ages of the various granitoid facies to be determined.

This section will address such relationships based on key field exposures.

➊ In Map C (centre of square A4) there is a well exposed glacial pavement displaying age relationships of most of the respective grantoids in the Galwollie Hill mapping area. In the centre of figure 6:20 there is a large composite raft comprised of Thorr Granodiorite, GWH 1 tonalite, pegmatite and micrograniticdykes. Around this large xenolith are further angular composite masses of the GWH 3 pink porphyritic monzogranite and pegmatite. The host granitoid to these xenoliths is the GWH 5 monzogranite. The disruption of the early xenoliths is slight and it is almost possible to piece the fragments back together like a jigsaw. The following conclusions can be derived from this exposure.

i) Diffuse masses of coarse pegmatite are younger than the Thorr Granodiorite, GWH 1 and GWH 3 granite but older than the GWH 5 monzogranite, as indicated by very sharp truncated contacts present along the exterior margins of the composite xenoliths.

ii) The microgranite dykes within the Thorr and GWH 1 xenoliths are not observed in the GWH 3 monzogranite, and are therefore older than GWH 3.

iii) All granitoid phases comprising the composite xenoliths were apparently highly competent during the intrusion of the GWH 5 monzogranite as indicated by sharp and angular contacts.

➋ Map C (lower half of C6):- the exposures here are excellent and display well the composite nature of the xenoliths within the Galwollie Hill area. In figure 6:21 a large raft, up to 20 metres long, of Thorr Granodiorite, GWH 1 tonalite and pegmatite have been broken up by GWH 5. The absence of pegmatite within this latter facies is particularly striking. At this same exposure there are abundant angular xenoliths of GWH 3 monzogranite, pegmatite and GWH 1 tonalite within the GWH 5 monzogranite. (figure 6:22).

➌ Figure 6:23 is 100 metres to the west of Map C and displays the age relationship between the GWH 1 tonalite and GWH 3 monzogranite. In this exposure angular planar xenoliths of GWH 1 have been fractured and intruded by GWH 3 which itself has been intruded by the GWH 5 monzogranite. It is also apparent from this location that there have been several phases of pegmatite intrusion due to the GWH 3 monzogranite truncating pegmatite intruded into the GWH 1 tonalite. As observed in ➊ there is another major phase of pegmatite that is clearly younger than GWH 3. It is therefore apparent that there has been a protracted history of pegmatite emplacement within this area.

Figure 6:20:- angular composite xenoliths of Thorr Granodiorite, GWH 1 tonalite, pegmatite, microgranitoids and GWH 3 monzogranite which have been "shattered" by the GWH 5 monzogranite. N.B it is possible to piece these xenoliths back together.

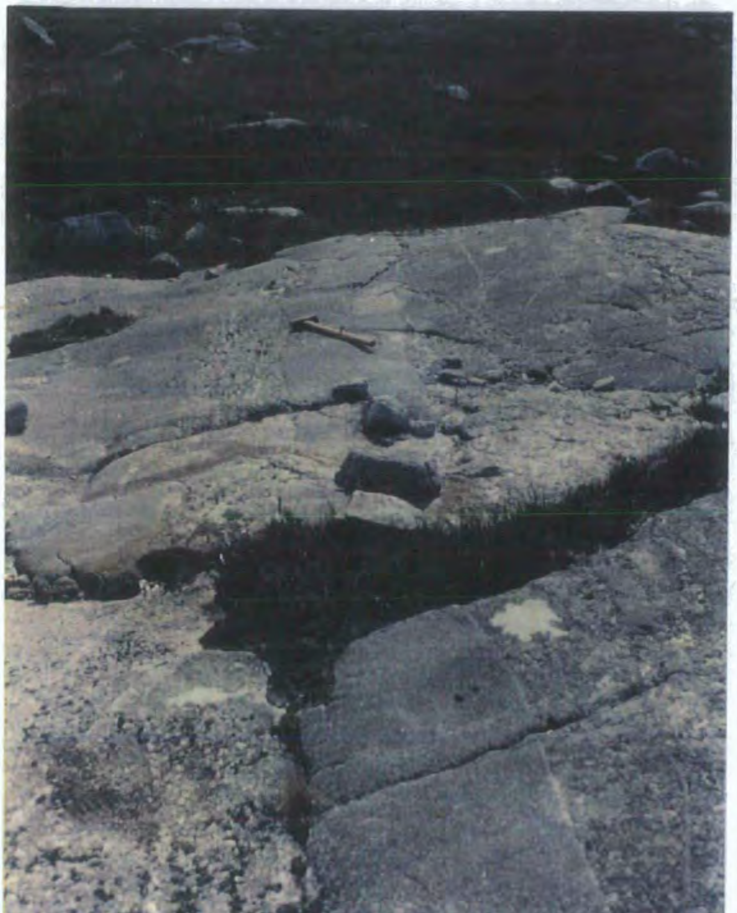


Figure 6:21

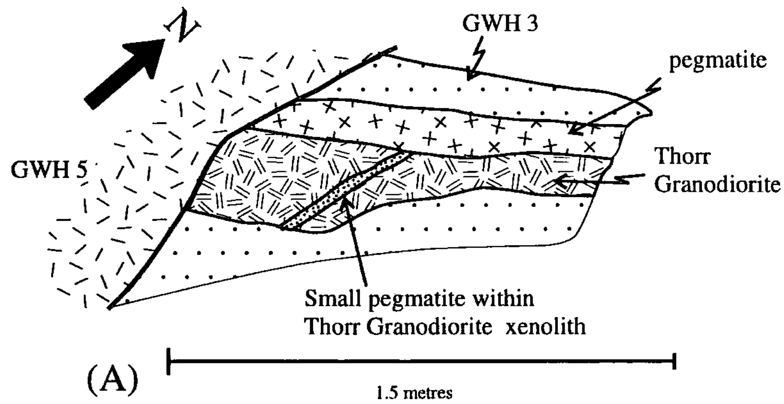
Large composite granitoid rafts composed of Thorr Granodiorite, GWH 1 tonalite and pegmatite which have been split apart by the GWH 5 monzogranite.

Figure 6:22:- Angular composite xenolith of GWH 2, GWH 3, microgranitoid and pegmatite within a host of heavily foliated GWH 5. These relationships suggest earlier granitoids were highly competent during the intrusion of GWH 5.

Figure 6:23:- large angular blocks of GWH 2 which have been intruded first by the GWH 3 monzogranite and then by the GWH 5 monzogranite (see text).

❹ Age relationships between the GWH 1 tonalite and Thorr granodiorite are less common due to the minor presence of these granitoids in this area. In Map C (centre of square C5) a vein of pegmatite within Thorr is truncated by both the GWH 1 and a later phase of pegmatite (detail of exposure is to small to be marked in the map clearly). In other areas xenoliths of Thorr can be seen totally surrounded by GWH 1 (e.g. Map C: upper H4) (figure 6:24a).

A similar relationship exists between the GWH 2 monzogranite and Thorr Granodiorite with xenoliths of the latter found within the former (e.g. Map C upper G5).



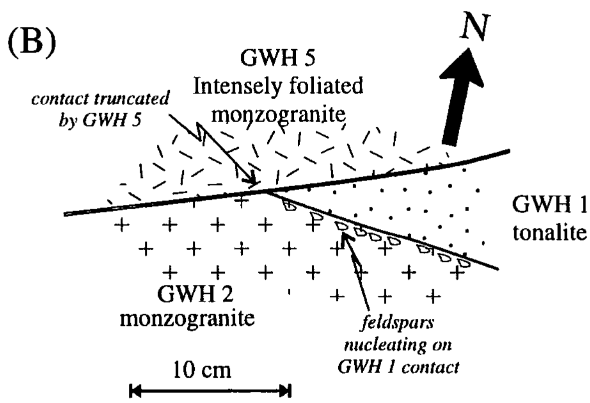


Figure 6:24:- More subtle cross-cutting relationships in the Galwollie Hill area. a) Relationship of GWH 1 tonalite to Thorr Granodiorite where pegmatites within the latter are truncated by the former. b) GWH 1 - GWH 2:- feldspars of GWH 2 growing on the boundary of GWH 2 implying the former nucleated on the latter and is therefore younger.

❺ The age relationship between the GWH 1 and GWH 2 granitoids is generally more subtle. Throughout the mapping area these granitoids tend to be commonly in contact with one another. Within square H5 GWH 1 xenoliths are seen enclosed within the GWH 2 tonalite. Figure 6:24b shows younging along the contacts of these two granitoids where there are coarse feldspar crystals within the GWH 2 monzogranite nucleated on to the sharp planar contact of GWH 1. (N.B.:- note the sharp truncation of the GWH 1-GWH 2 contact by the GWH 5 monzogranite).

❻ No direct age relationship was observed between the GWH 2 and GWH 3 granites, although the overall outcrop relationships suggest GWH 2 is older, i.e. GWH 2 monzogranite is more commonly seen in contact with GWH 1 than is GWH 3.

6:4:3:3 Deformation within the Galwollie Hill area

Outcrop

Deformation within this area is overall quite heterogeneous in character tending to be more intense within the younger granitoid phases. The earlier xenoliths tend to be only weakly deformed with only moderately developed S_6 foliation. The younger granitoids, especially the GWH 5 monzogranite, has well developed C' shear band cleavage (see figure 6:25).

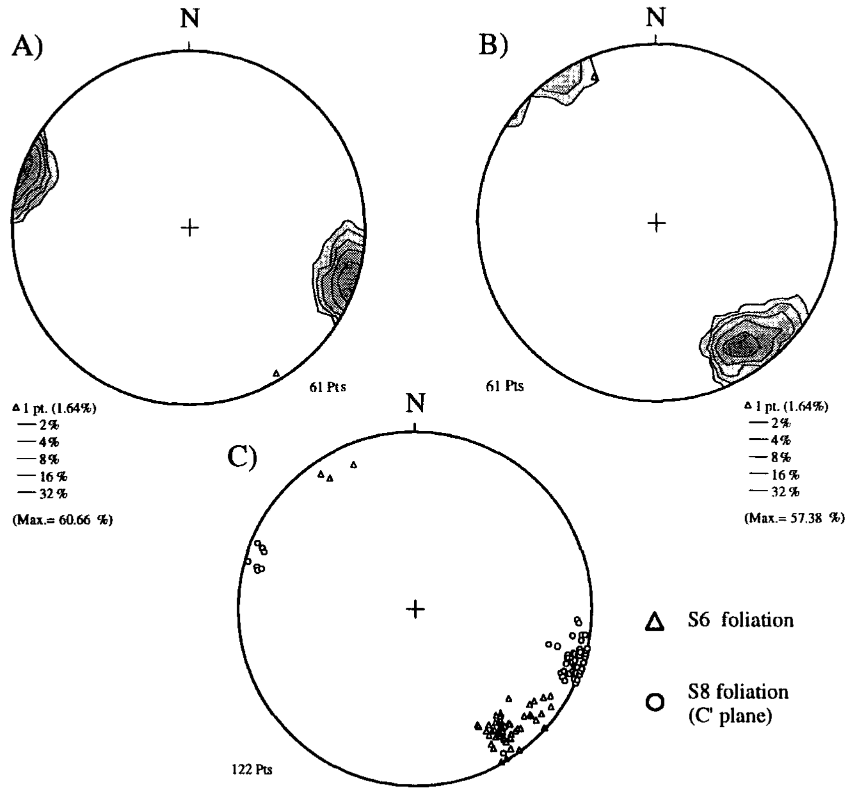


Figure 6:25:- Stereonet of the S_8 and S_6 foliation within the Galwollie Hill mapping area a) Contoured S_8 plot. b) contoured S_6 plot. (c) Composite plot of S (S_8) and C' (S_8) fabrics.

At the outcrop scale the youngest facies in Galwollie Hill is the most heavily deformed in contrast to the weakly deformed composite xenoliths. This leads to the conclusion that the GWH 5 monzogranite was being actively deformed as it was cooling. In figure 6:26 there is an example of where the foliation is partially deflected around a xenolith of earlier GWH 2 tonalite. The foliation in GWH 5 appears to stop against the tonalite xenolith with the latter only showing a very weak foliation. Around this xenolith the most intense deformation was observed 10 cm away from the contact. Furthermore the xenolith appears to be deforming in a brittle fashion (see quartz vein in fig 6:26) whilst the GWH 5 was deforming in a ductile manner as indicated by the deflection of the foliation near this bend. These relationships strongly suggest the host granitoid (GWH 5) was accommodating most of the strain during its subsequent emplacement. The maximum intensity of the foliation at 10 cm away from the contact may imply some form of viscous drag. This would account for the weakly deformed appearance of the xenoliths. Solid-state strain may have continued to be localised into this monzogranite as it cooled and became progressively overprinted with lower temperature deformation fabrics.

Thin Section

In this section the microscopic deformation fabrics will be compared between the less deformed xenoliths and the more highly deformed GWH 5 monzogranite seen in outcrop.

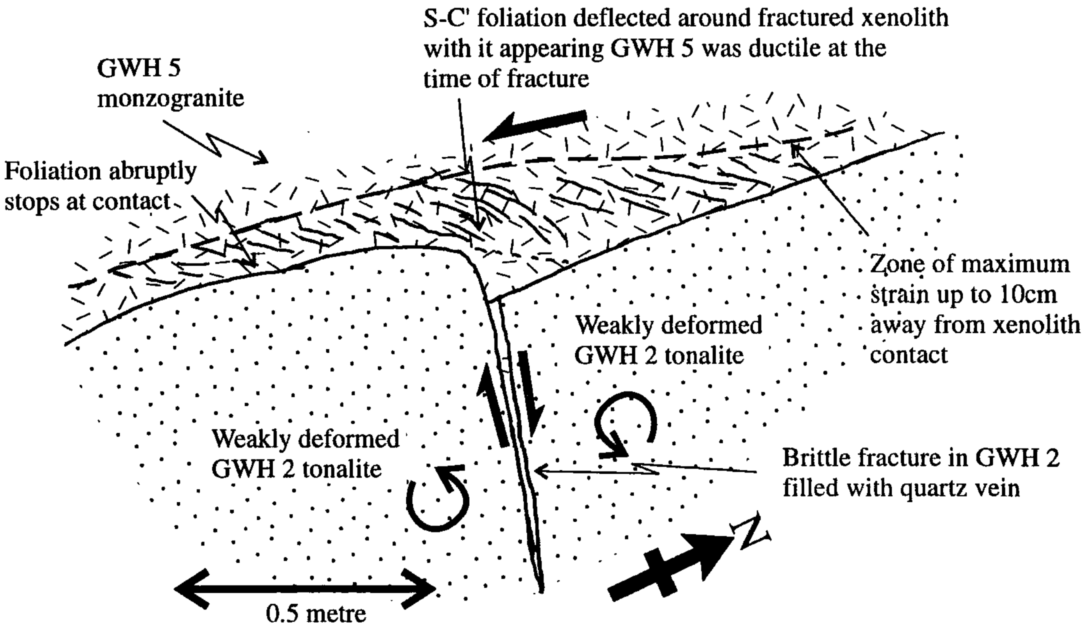


Figure 6:26:- sketch of the intense deformation within the GWH 5 monzogranite around a weakly deformed tonalite (GWH 2) xenolith.

Deformation within the GWH 1 and GWH 2 granitoids

In outcrop no C' shear band cleavages were observed within these granitoids. Plagioclases which commonly have euhedral crystal forms show very faint PFC fabrics. The preferred orientation of these laths is parallel to the weak foliation as indicated by aligned biotites. Within these plagioclase laths there are generally no signs of internal deformation, although in some samples some of the plagioclases have been bent and subsequently show "sweeping" undulose extinction plus some microcracking or kinking. Quartz displays the most evidence of deformation within these rocks with it deforming consistently in a ductile fashion. Most of the quartz grains are highly strained with well developed undulose extinction. Within large quartz grains subgrains are well developed as well as deformation lamellae, these features indicating that dislocation climb has occurred within the quartz grains of these granitoids. Generally these granitoids have only undergone limited dynamic recrystallisation by SR and GBM. These new grains are much finer grained than the highly strained quartz grains which form the matrix to this tonalite. Biotite (variably chloritised) forms the main foliation within these granitoids and shows minor kinking, especially where in close vicinity to feldspar laths. The GWH 2 monzogranite shows similar microstructural features although microcline is present. Along the margins of this mineral myrmekite is developed and within some of these crystals flame perthite is also seen. Generally these earlier granitoids have been deformed at medium-temperatures (although still well below the brittle-ductile transition).

Deformation within the GWH 3 and GWH 5 monzogranites

The GWH 3 monzogranite is variably deformed within the mapping area with the C' shear band S_8 cleavage not always developed. The GWH 5 monzogranite almost always contains a well developed S_8 cleavage. The dextral antithetic set is rare to absent in this area.

Plagioclases within these facies generally show greater evidence of internal deformation such as kinking with sweeping undulose extinction developed. Any PFC fabrics within these granitoids would have been obliterated by intense CPS modification. Microcline has been intensely myrmekitised with fine-grained masses of quartz and albitic plagioclase surrounding these reduced microcline porphyroclasts. Within the feldspars there is no evidence of any ductile deformation or recrystallisation. Two main types of deformed quartz occur within these monzogranites:- (a) large grains which display strong undulose extinction. These grains have been extensively flattened into the foliation. It is this strong flattening which gives this granite a highly deformed appearance on weathered surfaces.

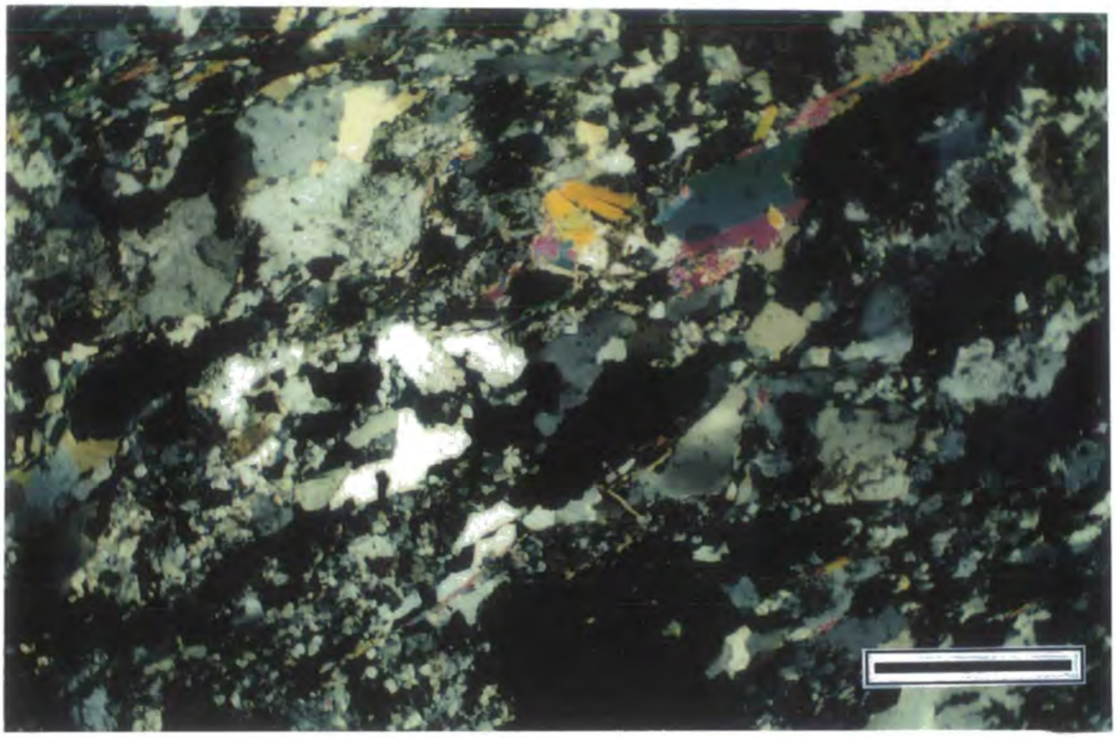


Figure 6:27a:- photomicrograph of highly deformed GWH 5 monzogranite. Extensive grain size reduction in quartz indicating dynamic recrystallisation. Biotites define sinistral S-C' fabrics (scale bar = 1mm).

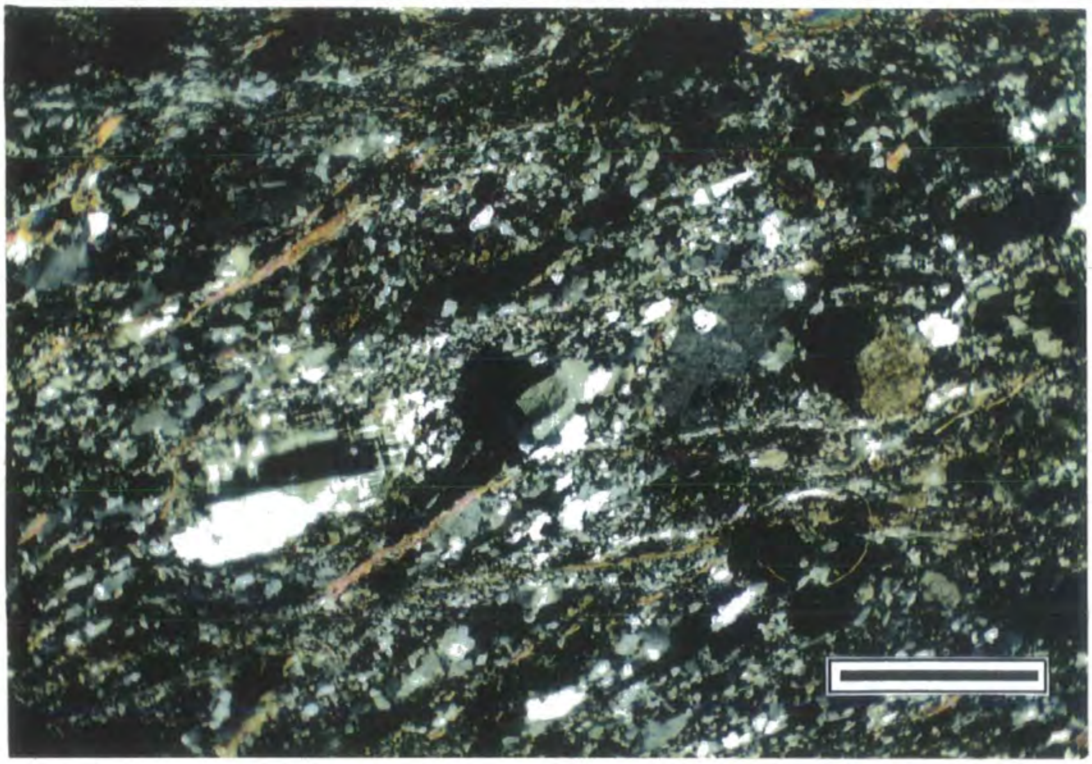


Figure 6:27b:- photomicrograph of mylonitised GWH 5 monzogranite. Quartz forms very fine-grained strain-free aggregates (recrystallised) whilst feldspar is reduced in size by either cataclasis or dynamic recrystallisation along its margins (scale bar = 1mm).

Within most of these deformed quartz grains deformation lamellae and subgrains are extensively developed. (b) The finer grained dynamically recrystallised quartz is generally much finer grained than the larger deformed quartz grains in (a) and most commonly lies along the shear band cleavage (C'~S_8). Around feldspar quartz is very fine-grained implying it is more highly strained. Passchier & Trouw (1996) state that "stiff" porphyroclasts will promote higher stresses to develop around them. Both GBM and SR recrystallisation mechanisms have occurred within the quartz grains of the GWH 5 monzogranite. The micas within this facies are strongly aligned along the S_6 foliation and within the S_8 shear band foliation. Some of these micas have been bent but the majority appear undeformed suggesting regrowth by pressure solution mechanisms. In some areas of the section muscovite forms "mica-fish" analogous to the Type II S-C mylonites of Lister & Snoke (1984). There appear to be two phases of muscovite, the first of which is seen in association with biotite. The second type appears to be later and post-dates the deformation forming undeformed grains up to 1.5 mm in diameter. In summary, the texture of the GWH 5 monzogranite (and to a lesser extent the GWH 3 monzogranite) is protomylonitic with the grain size of this granite having been considerably reduced (see figure 6:27a). In some localised zones intense granite mylonites have been produced (see figure 6:27b) Deformation is generally higher within these monzogranites than in earlier granitoid xenoliths on Galwollie Hill as indicated by the more extensive recovery of quartz, by dynamic recrystallisation.

6:4:3:4 Some Conclusions from Galwollie Hill

The following conclusions have been based on a comparison of each of the granitoid facies on Galwollie Hill with granitoids in the nearby areas.

1 The GWH 4 and GWH 5 monzogranites are essentially the same with the former being a less deformed variety of the latter. The presence of muscovite within these two monzogranites supports their correlation, as muscovite-bearing facies are generally rare within the Main Donegal Granite. It is believed that these two monzogranites belong to the G II member of the Trawenagh Bay pluton which has dominantly intruded into a fairly competent Main Donegal Granite.

2 The GWH 5 (G II) granite is not seen everywhere in Galwollie Hill as there appears to be relatively large areas of dominantly GWH 3 monzogranite, GWH 2 monzogranite and GWH 1 tonalite plus large volumes of pegmatite with occasional Thorr Granodiorite. In general there are, within this area, larger preserved masses (xenolithic or autochthonous) of pre-GWH 4 & 5 granitoids. In Map C these earlier granitoids have been extensively broken up, to form relatively small composite rafts, by the GWH 5 monzogranite.

3 Deformation is generally quite heterogeneous in this area with the majority of the deformation confined to the GWH 5 monzogranite and to a lesser extent the GWH 3 monzogranite. One might argue that the more deformed nature of these facies is due to the higher percentage of quartz, a mineral highly susceptible to ductile deformation. The microstructural evidence already described for all the mineral phases within the respective granitoids is consistent with the GWH 5 monzogranite being more intensely deformed with this granitoid and having undergone more extensive dynamic recrystallisation. It is the author's belief that this monzogranite was being actively deformed during its emplacement. The partitioning of strain into this less competent monzogranite is apparent from the deflection of the foliation within the granite around xenoliths of earlier formation (figure 6:26). In some exposures the foliation is oblique to the xenolith contacts and often there is a decrease in the intensity of the foliation away from the xenoliths. This would therefore suggest that all the earlier granitoids of this area were not subjected to such intense deformation during their emplacement as if this was the case one would expect the xenoliths to be the most highly deformed whilst the host would be the least deformed. This is definitely not the case in Galwollie Hill.

4 The granitoids of Galwollie Hill are generally quite heterogeneous in character and display evidence of considerable disruption, hence the name "the Galwollie Hill Zone of Disruption" (GHZOD). To the immediate NW of the summit of Galwollie Hill the granitoids become much more homogeneous with the number of large composite rafts or xenoliths drastically decreasing. This "zone of disruption" can be traced in a north-easterly direction and can clearly be observed on the lower SW slopes of Croaghleconnel, although the granitoids resembling the GWH 5 are not seen. The GHZOD lies along strike from the Doocharry Synform where there is well developed granitic banding which is believed to have formed by deformation of granitoids around raft-like masses of finer grained granodiorite (Pitcher & Berger 1972). There may be a possibility that these areas are connected, i.e. zones of high syn-magmatic strains.

5 The regularly banded granitoids within the Doocharry area consist of mafic-rich tonalites (the dark bands) and porphyritic monzogranites (the light bands). In the Galwollie Hill area there is no regular banding present. It may be that this banding has been broken up by the GWH 3 monzogranite, the pegmatite and the GWH 4 & 5 monzogranites. Modal comparisons of the GWH 1 and GWH 2 granitoids in Galwollie Hill with the dark and light bands in Doocharry, performed by Berger (1967;1971), show a very strong similarity. Therefore the

GWH 1 and GWH 2 granitoids may be the highly disrupted remnants of the banded granitoids which are seen in their entirety at Doocharry.

6 The significance of the GWH 3 monzogranite is uncertain from the mapping area alone i.e. is it an earlier granitic pulse belonging to the Trawenagh Bay Granite or is it related to the Main Donegal Granite?. It was compared with the pink porphyritic monzogranites of the NW half of the Main Donegal Granite. Despite both monzogranites being porphyritic the texture of these two granitoids is different, i.e. the GWH 3 monzogranite contains slightly less biotite and contains more megacrysts of plagioclase and K-feldspar. The re-mapping of Trawenagh Bay Granite by Pitcher may shed light on the significance of GWH 3. If the junction of G I and G II is projected into the Galwollie Hill (see figure 6:17 or Map D) area Map C lies somewhere long it. Therefore the GWH 3 monzogranite may correspond to the G I of the Trawenagh Bay Granite which has also broken up the earlier components of the Main Donegal Granite (i.e. GWH 1 and GWH 2) but has been broken up itself by the muscovite-biotite GWH 4 & 5 which belongs to the G II pulse of the Trawenagh Bay Granite.

7 Within the Galwollie Hill area there has been a protracted history of pegmatite emplacement with large volumes still preserved. Pegmatites were emplaced throughout the intrusion of the main granitoid phases and outlasted it with the occurrence of small pegmatite veins within the GWH 5 monzogranite. By far the most voluminous episode of pegmatite emplacement in Galwollie Hill was that post-dating GWH 3 (and by inference the earlier phases) but pre-dating GWH 4 and GWH 5. The significance of this pegmatite in relation to the Main Donegal and Trawenagh Bay Granites will be discussed in more detail in the following section.

6:4:4 The relationship of the Trawenagh Bay Granite to the Main Donegal Granite

The re-mapping of the Trawenagh Bay Granite by Pitcher and the detailed mapping by this author in the Galwollie Hill area has greatly improved the understanding of the relationship between these two plutons. The field evidence strongly suggests the Main Donegal Granite is older than the Trawenagh Bay Granite. At Meenderryherk the G I sheet can be observed intruding into the Main Donegal Granite immediately to the north of the Meenderryherk Septum.

At Galwollie the relationship of the plutons is much more spectacular with both the G I and G II pulses of Trawenagh Bay pluton intruding into this area. The original meta-sediments of this area (of which little is now left) have been invaded by three plutons with many of the relationships of each of the plutons still preserved. The presence of highly angular contacts between the respective granitoid phases implies they were all competent during the progressive intrusion of subsequent granitoid phases. From the Galwollie Hill area one can also see that the G I (GWH 3) member was highly competent during the intrusion of the more muscovitic G II (GWH 4 & GWH 5) member with sharp, often angular contacts, commonly preserved. Within the Trawenagh Bay Granite itself the outcrop patterns of G II relative to G I also supports the hypothesis that G I was relatively competent also during the emplacement of G II. The major phase of pegmatite intrusion in the Croaghleconnel-Galwollie Hill appears to be post-G I but pre-G II. This relationship is well seen in Map C where the G II monzogranite is seen relatively devoid of pegmatitic intrusions apart from the later more regular pegmatite veins. Pitcher & Read (1959) originally stated that the pegmatite fringe extended from Brockagh to the Galwollie Hill area. In the Brockagh area the pegmatite is mainly confined to the marginal facies and was not seen in the MDG pink porphyritic monzogranites to the SW which implies these latter porphyritic monzogranites are younger. Similarly it was also not observed in the G I phase to the north of the Meenderryherk Septum-Brockagh marginal facies which again implies G I is younger than the pegmatite of this area. This relationship is in contrast to Galwollie Hill area where the G I has been extensively intruded by pegmatite. This therefore implies the voluminous phase of pegmatites in the Croaghleconnel-Galwollie Hill area are *younger* than the marginal facies pegmatites seen within the Brockagh area. This further supports the evidence that there was a major phase of pegmatite intrusion during the emplacement of the Trawenagh Bay pluton (post G I- pre G II). On the whole the majority of this pluton is relatively devoid of pegmatite which implies there may have been some structural control on its greater abundance in the Galwollie Hill area, i.e. it may have ascended the north-western boundary of the MDGSZ.

The full extent of the intrusion of the G I and G II pulses into the MDG is not certain and may require some additional mapping of the area. The G I monzogranite was observed at GR B 835042, some 600 metres to the NE of Galwollie Hill summit but there were no granitoids resembling this seen any further to the NE. Beyond the poorly exposed ground lying within the Glenaltaderry, the degree of exposure is quite good on the south-eastern flanks of Croaghleconnel. Here, the granitoids are quite heterogeneous with the dominant granitic phase resembling the G I (GWH 3) monzogranite. This monzogranite can be observed from the bend in the old track (GR B 845049) which goes to Derryleconnel Near. The north-easterly extent of this G I phase is not certain. An analogy of the outcrop relationships in the Galwollie Hill area is like an imaginary "sea" of G II monzogranite in which there are icebergs of older Thorr Granodiorite, Main Donegal facies and G I (Trawenagh Bay) floating.

The Main Donegal Granite can be envisaged as a ice sheet which has been progressively broken up by Trawenagh Bay Pluton. From this analogy it can be understood why the supposed boundary between these two plutons was so elusive to Pitcher & Read (1959).

It is this authors belief that the Trawenagh Bay Granite and the Main Donegal Granite are separate plutons. Compositionally and petrographically these plutons are very similar but there does appear to be a number of subtle differences:-

1) The granites of the Trawenagh Bay pluton are relatively rich in muscovite. This feature has not been seen by the present author within any of the major granitoid phases of the Main Donegal Granite, although the marginal sheets of this pluton are muscovite bearing.

2) The textures of the Trawenagh Bay Granites are slightly different to those of the Main Donegal Granite. The G I and G II phases are typically medium to coarse-grained, relatively equigranular to weakly porphyritic monzogranites. The petrographic equivalent monzogranites in the Main Donegal pluton are texturally different tending to be porphyritic with large megacrysts of microcline in a relatively fine groundmass.

In regards to deformation it appears that the Trawenagh Bay pluton was intruding across the boundary of the MDGSZ as G I and especially G II are the most intensely deformed. The deformation within the Galwollie Hill area is very heterogeneous with the majority of the strain localised into the later granitoid phases with the earlier phases being only weakly deformed. The reason for the earlier granitoid phases being less deformed than the latter is uncertain as one would expect to see a similar relationship to the Carbat Gap where the oldest xenoliths are the most deformed. What is clear is that the western part of the Trawenagh Bay Granite was syn-kinematic in relationship to the Main Donegal Granite Shear Zone.

6:5 Summary

The chapter has addressed the relationship of the Main Donegal Granite to older and younger adjacent members of the Donegal Batholith. The most clear evidence for the MDGSZ being in operation prior to the emplacement of the Main Donegal Granite are the tonalite xenoliths of the Carbat Gap. These xenoliths have been shown to dominantly belong to the Ardara pluton and tend to be more intensely deformed than the immediately adjacent MDG facies. The preservation of tonalite country rock contacts implies they are dominantly in situ and have therefore been within the shear zone for longer. The Main Donegal Granite has been emplaced at a later stage and therefore has not witnessed the same degree of deformation as that seen in the tonalite xenoliths. This relationship implies that the intrusion of the Ardara Pluton was syn-kinematic in relation to the Main Donegal Granite Shear Zone.

The relationship of the Thorr pluton to the MDGSZ is less certain as there is less good evidence for syn-kinematic relationships. The truncation of the F_6 folds (figure 6:5) by the Thorr pluton would imply the shear zone was present before the emplacement of the Thorr pluton, or at least the prolongation. McErlean (1993) reports the presence of magmatic fabrics attributable to the MDGSZ in the Crohevy Hills which in a southwards direction become progressively overprinted by solid-state fabrics. Therefore McErlean (1993) believed that the southern part of the Thorr pluton was syn-kinematic in respect to the MDGSZ. The presence of the Thorr "prolongation" in the present area of the Main Donegal Granite is possibly the strongest evidence for shear zone controlled emplacement with the southern part of this pluton having exploited a major NE-SW trending weakness in the crust of NW Donegal.

The Trawenagh Bay pluton, from the field evidence, has been shown to be dominantly younger than the Main Donegal pluton. In the Galwollie Hill area the G II facies is the most intensely deformed phase present and implies the MDGSZ was still in operation after the emplacement of the Main Donegal Granite. The presence in the Galwollie Hill area of earlier xenoliths that are less deformed than later granitoids implies the deformation within the shear zone has been quite complex with the distribution of strain throughout the pluton being quite heterogeneous in nature. An alternative explanation is that movement along the MDGSZ has been episodic with parts of the Main Donegal pluton having been emplaced in periods of possible quiescence or periods of reduced displacements. These factors will be addressed in the final chapter which is concerned with the evolution and construction of the Main Donegal Granite.

Chapter 7

The petrography and geochemistry of the Main Donegal Granite

7:1 Introduction

In this chapter the petrographic and geochemical variation within the granitoids that comprise the Main Donegal Pluton will be discussed. The initial aim of this sampling was to identify possible cryptic variations which may exist between similar looking granitoids and to also use geochemical fingerprinting to compare the different granitoid facies from the respective mapping areas within the pluton. These data have been used in alliance with the field evidence and petrographic studies (thin-sections) to understand the variation within the pluton. Most of the samples were obtained from the areas where detailed mapping has been performed, i.e. Maps A, B & C. In total the major and trace element compositions were obtained from 223 samples using the XRF analytical method, mainly from powdered briquettes (see Appendix D for preparation methods).

The main topics to be addressed in this chapter are as follows:-

i) A brief insight into the possible source regions to the Donegal granites, summarising the results of previous researchers.

ii) The petrographic variation within the granitoids of the Main Donegal Pluton.

iii) The overall geochemical variation within the Main Donegal Pluton.

iv) Geochemical variation within the pluton. This will be addressed in respect to the heterogeneous zones of Chapter 4 and the homogeneous zones of Chapter 5. In each of the two sections references will be made to the mapping areas within these zones.

7:2: Summary of published research into possible source regions to the Donegal Batholith

It is important to re-emphasise, at the beginning of this section, that this thesis does *not* include either mineralogical or geochemical studies aimed at a better understanding of the *genesis* of the Donegal Batholith magmas. Nevertheless, the specific topics of this chapter are best prefaced by a summary of published ideas of magma genesis.

Many authors have speculated that the basement rocks, and possible source, to the Donegal Granites are similar to the Lewisian Complex (Pitcher & Berger 1972; Bowes & Hopgood 1975 and Dempsey *et al.* 1990) which was believed to underlie parts of NW Donegal. This is due to the presence of lithologies resembling the Lewisian Complex outcropping to the north of Donegal, notably on the Rhinns of Islay and on the small island of Inishtrahull, just to the north of Malin Head (see figure 2:2 of this thesis for location). It was believed that the Dalradian, and also possibly underlying Moine, may have contaminated the Lewisian partial melts during the ascent of the Donegal Granite pulses.

More recent work questions the presence of the Lewisian basement to the south of the Great Glen Fault (e.g. Muir *et al.* 1994). Dating of the gneisses on Islay and on Inishtrahull by Marcantino *et al.* (1988) and Daly *et al.* (1991), respectively, by U-Pb zircon dating methods have shown a general concordance in age: this is approximately 1800 Ma, i.e. Proterozoic. Furthermore the mapping of this basement on Islay, Colonsay and on Inishtrahull (the Rhinns Complex) by Muir *et al.* (1994) showed that the geochemistry and petrography of the gneisses were dissimilar to the calc-alkaline tonalitic gneisses which typically form the bulk of the Lewisian Complex of NW Scotland and Outer Hebrides. The rocks of the Rhinns Complex are typified by syenitic gneisses which have been intruded by a suite of gabbroic sheets. Several phases of deformation have been identified and these are believed to be contemporaneous with the Laxfordian deformation recorded in the Lewisian Complex (Muir *et al.* 1994).

The Rhinns gneisses are believed to originate from mantle sources, as indicated by the absence of inherited Pb component within zircons and also by the small differences between their Nd model ages and the emplacement ages (Daly *et al.* 1991). These authors found no evidence to suggest the formation of the Proterozoic gneisses by reworking of Archaean (i.e. Lewisian) crust.

Some authors have speculated that this 1800 Ma Proterozoic basement may well underlie the Grampian Highlands of Scotland and the NW part of Ireland (Dickin & Bowes 1991: Muir *et al.* 1994). The extent has been inferred from indirect evidence, based mainly on the distribution of isotopic signatures (i.e. inherited zircons and Nd model ages) within the younger Caledonian Granites; these plutons generally lie between the Highland Boundary Fault and the Great Glen Fault. Many of these granite plutons do possess Proterozoic zircons and Proterozoic Nd model ages, which may well indicate the presence of such basement under them (Muir *et al.* 1994).

Within the granites of NW Ireland and mid-Scotland no inherited Archaean zircons were found, strongly implying the absence of the Lewisian Complex to the south of the Great Glen Fault (Dickin & Bowes 1991). Muir *et al.* (1994) expressed caution, stressing that many of these inherited Proterozoic signatures may have derived from incorporation of Moine or Dalradian rocks into these granite magmas (Muir *et al.* 1994; Dickin & Bowes 1991), which possess similar signatures due to possible erosion of a Proterozoic basement. Fitches *et al.* (1990) analysed igneous clasts from the Port Askaig Tillite, which are derived from the gneisses of the Rhinns Complex, implying that this basement was a sediment source during deposition of the Dalradian sediments. On mainland Donegal there is no exposed basement similar to that of the Rhinns Complex, and the Dalradian is believed to be in tectonic contact with the Proterozoic basement, separated by a zone of mylonites belonging to the Malin Head Thrust (Max & Long 1985).

The Donegal Granites show restricted range of initial Sr isotopes (0.7051-0.7068); values typical of the "newer" Caledonian Granites (Dempsey *et al.* 1990). These same authors stated that, nevertheless, the ε_{Nd} values were much more variable, ranging from -1.2 to -8.3. The combined restricted Sr isotope range and the more variable ε_{Nd} values "*preclude single crustal sources*" according to Dempsey *et al.* (1990). The most viable model that these authors present, to satisfy these isotope values, is as follows:-

- In crustal derived melts the abundance of Sr is quite low, due to plagioclase remaining as a residual phase in the source region. At mantle depths, in contrast, plagioclase is usually unstable, so that Sr behaves incompatibly and any mantle derived melts will be rich in Sr. According to Halliday *et al.* (1985), if this mantle melt mixes with lower crustal material, in the proportions 50%-50%, then the majority of Sr in the resultant magma (up to 90 %) may derive from the mantle. Applying these principles to the Donegal Granites, the low and restricted ranges of initial Sr ratios may be explained by dominance of mantle Sr, regardless of interaction between the rising melts and upper crustal material. In contrast the Nd concentration is not buffered in this way and hence *"during interaction with the crust highly variable Nd isotopic compositions ensue, dependent on the degree of contamination"* (Dempsey *et al.* 1990).

In the majority of the Donegal plutons there is some degree of zoning, in which younger pulses within individual plutons become more evolved. This relationship is still true for the Main Donegal Granite although it does not display an obvious zoned configuration. Harmon & Halliday (1980) attributed this trend to an upward melting progression in the depth of the melting process that generated the magmas, from mantle to lower crust with the earlier pulse being the most primitive.

The same authors stated that, in a two-component system, mixing models suggest the requirement of 30-80% of continental material to account for the observed Sr and O isotopic ratios.

For the Main Donegal Granite and Trawenagh Bay Granites, the more highly negative ε_{Nd} values (when compared to the other Donegal plutons) implies these granites have possibly undergone higher degrees of contamination by old basement, i.e. possibly the Rhinns Complex. O'Connor *et al.* (1982) investigated whether or not the Main Donegal Granite has witnessed much contamination at the present level of exposure, by plotting the calculated initial $^{87}Sr/^{86}Sr$ against present day $^{87}Rb/^{86}Sr$ ratios. Some of their samples did plot on a linear regression, implying that some degree of contamination from Dalradian meta-sediments may have occurred, but from overall field evidence it is believed that contamination within the pluton is minimal (Pitcher & Read 1959; and present findings). This feature will be considered in section 7:4:1:1).

Therefore the source region to the Main Donegal Granite may consist of a mantle component, which has been contaminated by lower crustal basement. The latter may resemble the Proterozoic Rhinns Complex and, to a lesser extent, there may well have been some degree of contamination from Dalradian metasediments at or just below the level of final emplacement. Relevant Sr-Nd isotopic data are shown in figure 7:1.

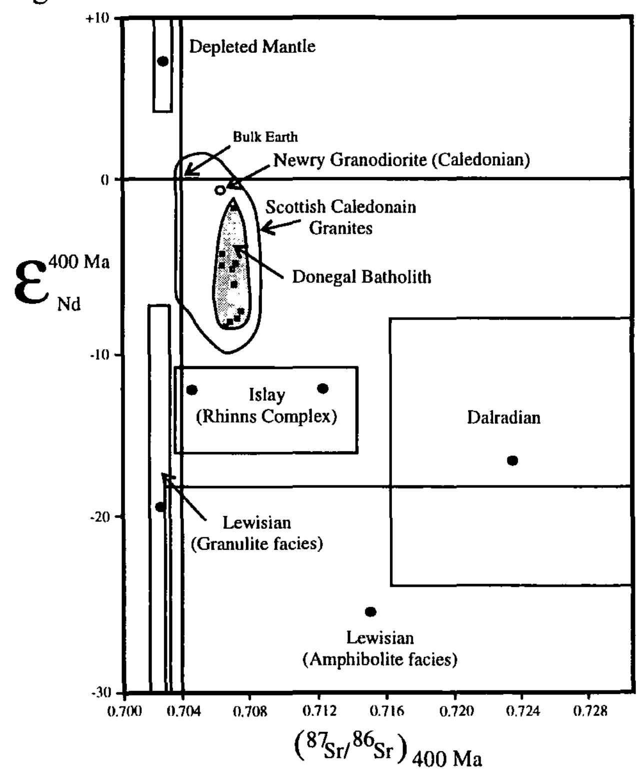


Figure 7:1
Possible source regions to the Donegal Batholith based on Nd and Sr isotopes. Modified from Dempsey *et al.* (1990). Rhinns Complex at 420 Ma from Dickin & Bowes 1991)

7:3 Petrographic variation within the Main Donegal Granite

In chapters 4, 5 & 6 the internal variation within the pluton was described, mainly on the basis of field evidence. This section will describe the overall petrographic nature of the granitoids. Up to this point in the thesis, thin-section studies have only been used to address the overall mineralogy and deformation structures within the granite. This section will describe igneous textures and will give an insight into the crystallisation orders of the mineral phases present. These mineral abundances within the different granitoid phases may then help to identify the causes of any geochemical variations. Figure 7:2 shows a IUGS-Streikeisen plot for the granitoids of the Main Donegal Granite, Trawenagh Bay Granite and xenolithic granitoids which belong to older plutons within the Donegal Batholith. See the Appendix C for modal analyses of each of the individual granitoids.

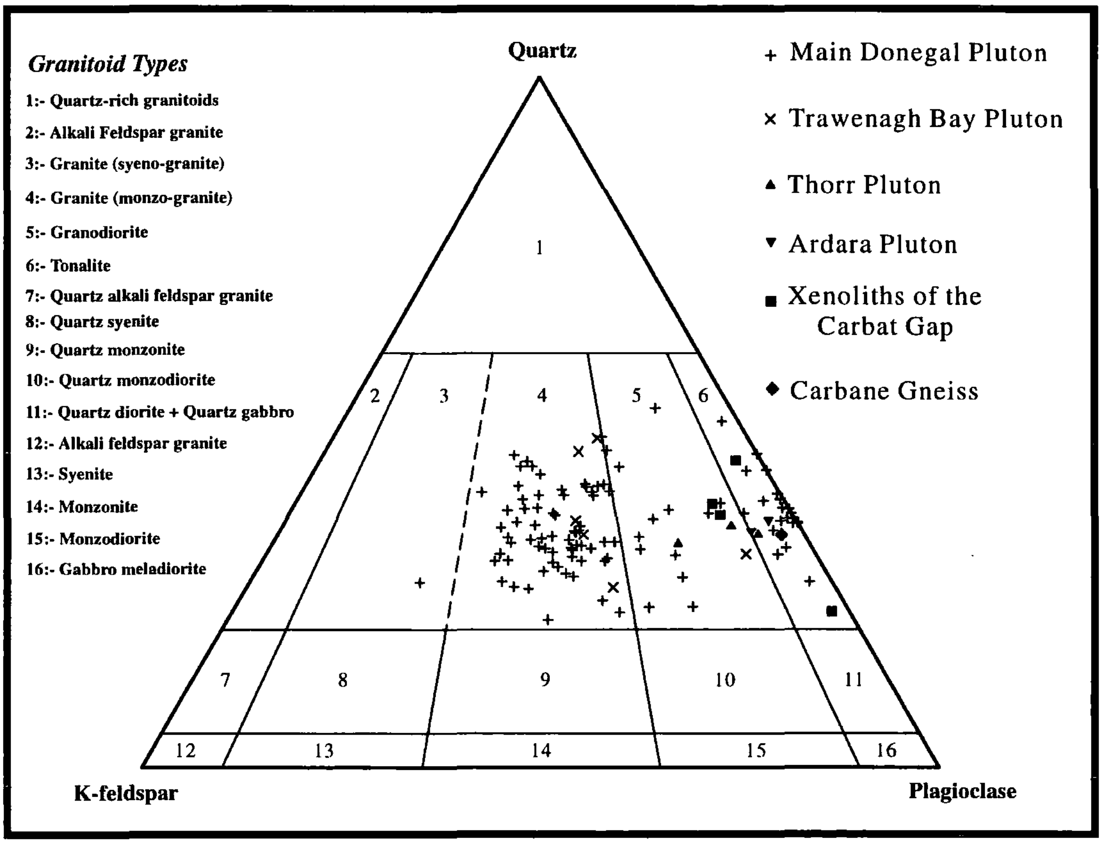


Figure 7:2:- IUGS-Streikeisen plot for the granitoids which occur in the Main Donegal and Trawenagh Bay plutons, plus older xenoliths within these plutons. (Also includes the xenoliths of older plutons). See appendix C for details of modal analyses.

Overall the composition of the Main Donegal Pluton ranges from equigranular, biotite tonalites-granodiorites to porphyritic (and to a lesser extent equigranular) monzogranites; the latter tending to be volumetrically the most abundant. The

Trawenagh Bay Pluton also tends to plot within the monzogranite field, although some xenolithic granodiorite material was encountered, although this is believed to belong to the Main Donegal pluton. Xenoliths of older plutons, i.e. Ardara, Thorr and Carbane Gneiss plot within the granodiorite and tonalite field and are distinguished from the tonalites of the MDG by their much higher mafic contents.

The majority of this section will describe the main granitoid types observed within the central areas of the pluton, as the almost entire petrographic variation is encountered in this area i.e. granodiorites, tonalites and porphyritic monzogranites. Furthermore, the effects of solid-state deformation is at a minimum in these areas, with igneous fabrics preserved. The more homogeneous monzogranites, which are situated closer to the margins, tend to be considerably more effected by solid-state deformation and hence will be discussed in less detail. The granitoids will be discussed in order of increasing K-feldspar and hence the tonalites will be described first.

7:3:1: The petrography of the tonalites (SRU 2, GM 2, CBH 4* & GWH 1)

The tonalites of the Sruhanavarnis Valley and Glendowan Mountains show the greatest degree of similarity in regard to texture and grain size, although the CBH 4* tonalite is more finer grained. As was mentioned in Chapter 4 the exact age significance of the Crobane Hill, CBH 4* tonalite, is uncertain and hence this tonalite may be of different origin to the tonalites of the other areas. Figure 7:3 (a & b) are photomicrographs of the tonalites which are typically seen within the MDG.

The most abundant mineral phase in the tonalites is plagioclase, which shows a wide variation in shape and size. Wherever possible, compositions of plagioclase have been obtained, using the albite twin extinction method of Michel-Levy. In most sections there were a few plagioclases positioned in the required orientation, i.e. normal to the (010) composition plane; in sections where no suitable crystals were found, the composition was estimated from the maximum extinction angle. The composition of plagioclase within the tonalites ranges from An_{27-36}, although around some plagioclases more albitic rims are occasionally seen, with An content <10%. On average the plagioclases tend to be no larger than 5 mm in diameter, with the larger crystals also tending to be the most euhedral in form. Furthermore it is these larger plagioclases which tend to be zoned, often oscillatory. The zoned crystals tend to show Carslbad twinning rather than albite twinning, although in some crystals there is patchy development of the latter. Generally more common are anhedral to subhedral plagioclases, which tend to be smaller (1-3 mm in diameter), twinned (albite and sometimes pericline) and often unzoned (see figure 7:3 a & b). Inclusions within plagioclase are generally quite rare but when encountered the most common mineral

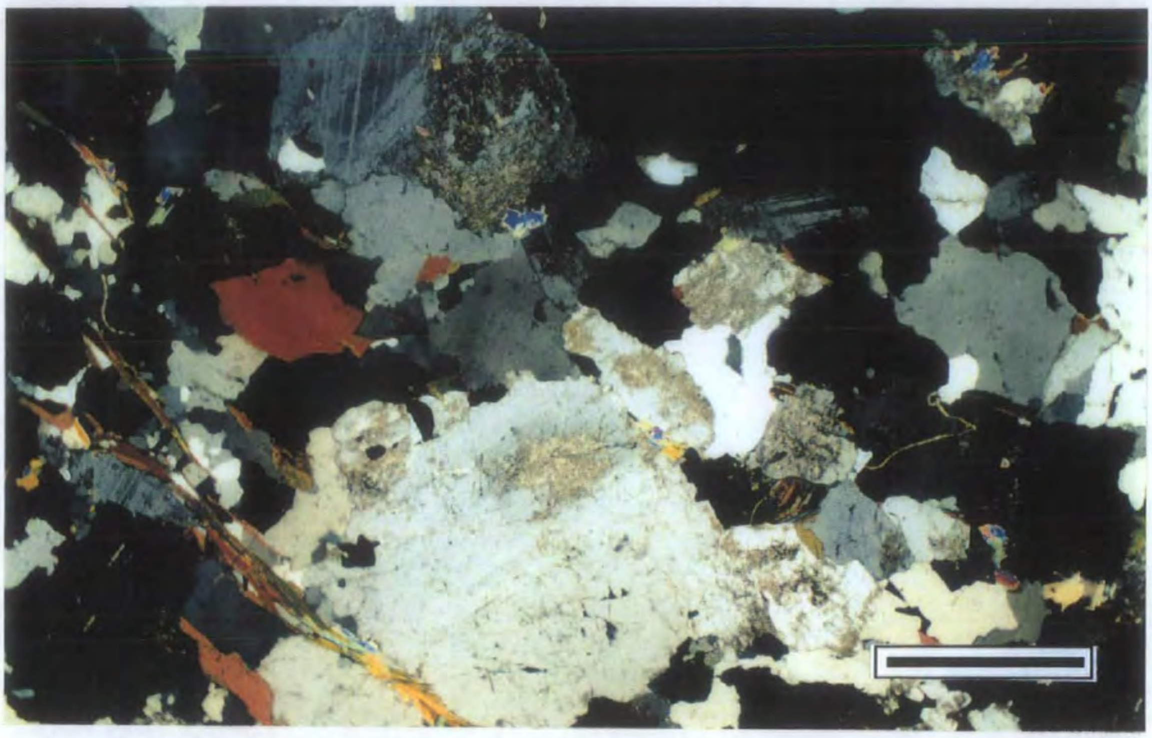


Figure 7:3a:- Photomicrograph of tonalite from the Glendowan Mountains (Map II) (red-brown biotite variety). The majority of the plagioclases tend to have an anhedral form with no zoning present. To a lesser extent there are larger more euhedral plagioclases which are commonly zoned and may represent a phenocryst phase (scale bar = 2 mm).

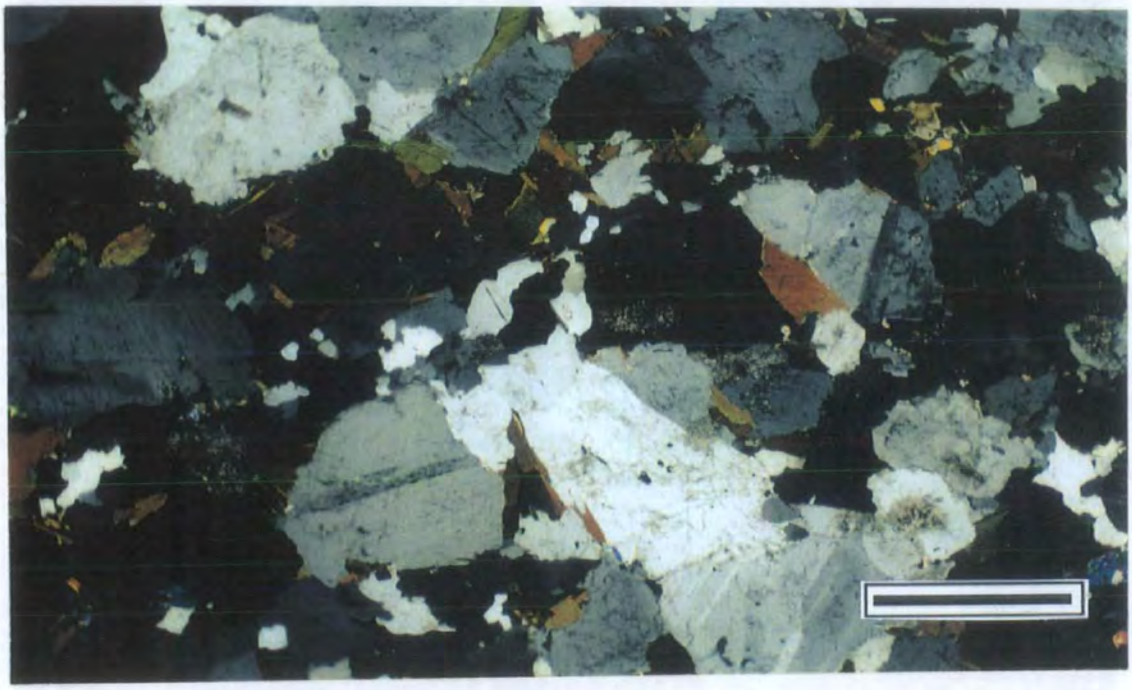


Figure 7:3b:- Photomicrograph of tonalite from the Sruhanavarnis Valley (Map B) (olive-green biotite variety). Biotite tends to occur around the margins of the plagioclase laths suggesting these laths may have served as nucleation surfaces during its crystallisation (scale bar =2mm).

inclusion is biotite and sometimes quartz, plus occasional accessory mineral phases. The overall textural relationships imply that plagioclase was dominantly the early phase crystallising within the tonalitic magma. The presence of biotite inclusions implies that there may have been overlap with biotite crystallisation.

Microcline is generally absent from the tonalites. In samples where it is present, it is generally interstitial and has a highly irregular crystal form. Within these tonalites the microcline is generally devoid of inclusions, in contrast with the larger, inclusion-rich microclines of the porphyritic monzogranite. In the tonalites it appears that the K-feldspar was a relatively late phase, possibly crystallising with quartz.

Apart from its overall distribution there is little to say about quartz because these crystals are generally highly strained due to later solid-state deformation which has already been addressed in Chapters 4, 5 & 6. In the central regions of the pluton quartz is only weakly deformed and therefore the present shape of quartz grains in these tonalites may be related to primary crystallisation (although slightly flattened due to crystal plastic strain). Quartz tends to form quite irregular grains which vary in size and shape. On the whole they contain no inclusions, although some large grains of quartz contain other phases such as plagioclase.

Biotite is relatively common in the tonalites, forming up to 15 % of the rock. In all of the tonalites studied the biotites tend to be in the highest concentration around crystals of plagioclase, which implies that the crystal edges of the latter may have acted as nucleating sites during the growth of biotite. In the Sruhanavarnis Valley and the Glendowan Mountains, the tonalite biotites varied in colour. In Map I of the Croaghacullin area the biotites in the GM 2 tonalite were olive-green, whilst in Map II of the Moylenanav area the biotites in GM 2 were red-brown. A similar situation was observed in the Sruhanavarnis area where, at the southern end of Map B, the SRU 2 biotites were red-brown but towards the northern end of this mapping area the olive-green variety was far more prevalent. The difference in colour is due higher proportions of Ti^{4+} and lower Fe^{3+} in the red biotites, compared with the olive-green variety (Deer *et al.* 1966; Berger 1967) . This different proportion of iron and titanium would be a primary feature of crystallisation and it therefore suggests the presence of cryptic variation within the tonalites of the heterogeneous zones. It was noted that the red-brown variety of biotite contains more inclusions of zircon than the olive-green variant. The degree of alteration within the biotites tends to be quite minimal, with only partial development of chlorite, mostly along the (001) cleavage plane. It was noted, that when even small amounts of microcline are present, the degree of chloritisation is generally greater. The relationship between K-feldspar abundance and biotitic chloritisation is particularly clear in the porphyritic monzogranites, as Berger (1967) showed. Within some of the tonalite there appears to be two phases of biotite growth, although whether or not this feature is due to magmatic or deformational (or both) processes is uncertain. "Fresh" biotite is commonly seen overgrowing earlier more altered biotite and is often aligned at an inclined angle to the earlier biotite. It is the early biotite which mainly defines the foliation within the tonalites. This fresh biotite is marked by planar crystal edges and "clean" pleochroism in contrast to the earlier biotite which tends to have irregular boundaries and very indistinct pleochroism. The irregular boundaries of this earlier biotite may be the result of pressure solution, with the new biotite growing in areas where stresses are lower. The lesser abundance of zircon inclusions in the later biotites would conform with the dissolution hypothesis. Dynamic recrystallisation in biotites tends to occur at higher temperatures (450-500°C) (Passchier & Trouw 1996) and is therefore probably not the cause of the second phase of "fresh" biotite.

Muscovite was seen in all tonalites of the central heterogeneous zone, although modal proportions were generally less than 1.5%. It occurs in two main forms:- i) From the alteration of plagioclase. Commonly muscovite plates tend to be randomly oriented in the central regions of intensely seriticised plagioclases. This muscovite appears to be of secondary origin. ii) The more common type is seen intergrown with biotite and commonly contributes to the foliation. Where seen in association with biotite, some of the muscovite clearly cross-cuts biotite, implying it is later. In other parts of sections the muscovite appears to have progressively replaced biotite as faint remnant patches of biotite are sometimes seen within muscovite crystals.

Accessory minerals are abundant in the tonalites, often forming inclusions within biotite or otherwise growing nearby. The most common accessories in the tonalites of the Sruhanavarnis Valley are apatite and elongate zircons, the latter with well developed pleochroic haloes. Within the CBH 4* tonalite epidote forms the main accessory, with the cores of these crystals commonly made up of allanite. Where enclosed in biotite, this more allanitic core has led to the development of a pleochroic haloes due to its content of radionuclides.

7:3:2 Petrography of the Granodiorites (SRU 1, GM 1 and CBH 2)

The granodiorites in the Crobane Hill area are far more abundant than the more fine-grained granodiorites observed in the Sruhanavarnis Valley and the Glendowan Mountains. In all three areas the granodiorites occur predominantly as rafts within the younger porphyritic monzogranites. In comparison with the tonalites, the granodiorites of the central zones overall tend to show more variable grain size.

Plagioclase is the most abundant mineral phase within the granodiorites, on average comprising 50% of the mode. The composition of plagioclase is more variable within the three mapping areas: Crobane Hill (An_{26-36}), the Sruhanavarnis

Valley (An_{24-28}) and the Glendowan Mountains (An_{28-38}). Within all sections it was observed that the largest plagioclases were the most euhedral, ranging from 2-5 mm in length, although on average the size was 2-3 mm. The plagioclases in the granodiorites on Crobane Hill and in the Sruhanavarnis Valley are much more poorly zoned than the plagioclases within the GM 1 (see figure 7:4a). The smaller plagioclases within the granodiorites are more anhedral and more commonly twinned (albite and to a lesser extent pericline twinning). In the larger plagioclases the albite twin lamellae are much more patchy. The degree of alteration within the plagioclases is generally very minor, with only slight sericitisation which tends to affect only one set of the albite twin lamellae. Occasional antiperthitic patches of K-feldspar are seen within some large plagioclases tending to occur as small "blobs". Inclusions within plagioclase tend to be biotite and to a lesser extent quartz. In summary, the overall textural relationships of plagioclase are very similar to that described in the tonalite section above.

Within these granodiorites K-feldspar (perthitic orthoclase or microcline) is relatively abundant, comprising on average 20% of the rock, although in some samples the K-feldspar content can be as low as 10%. In all sections it has an highly irregular form, interstitial to plagioclase, biotite and to a lesser extent quartz. Some larger crystals (3 mm in diameter) tend to be poikilitic, containing inclusions of quartz, biotite and intensely sericitised plagioclase where in contact with K-feldspar. Myrmekite is well developed, sometimes at contacts with plagioclase, with "wart-like" protrusions of vermicular quartz and albitic plagioclase growing into microclines. In summary, the absence of any euhedral K-feldspar argues against it forming early in the crystallisation of the granodiorite; it appears to have formed after plagioclase, biotite and the bulk of quartz (see figure 7:4b).

Anhedral quartz, a major component of the granodiorites, is generally weakly strained, showing varying degrees of undulose extinction with no evidence of any extensive dynamic recrystallisation having taken place.

The biotite content of the granodiorites is 15-16% and this mineral generally has a weak to strong preferred alignment. In comparison with the tonalites, the granodiorites are generally more biotite-rich. In the Crobane Hill and Glendowan Mountains the biotite colour is olive-green whilst in the Sruhanavarnis Valley the red-brown variety is also present, again implying the possibility of internal variation within these granodiorites. In all three areas there is also evidence of two phases of biotite; later fresh biotite with planar boundaries and distinct pleochroism contrasts with earlier biotite which has indistinct pleochroism and irregular boundaries.

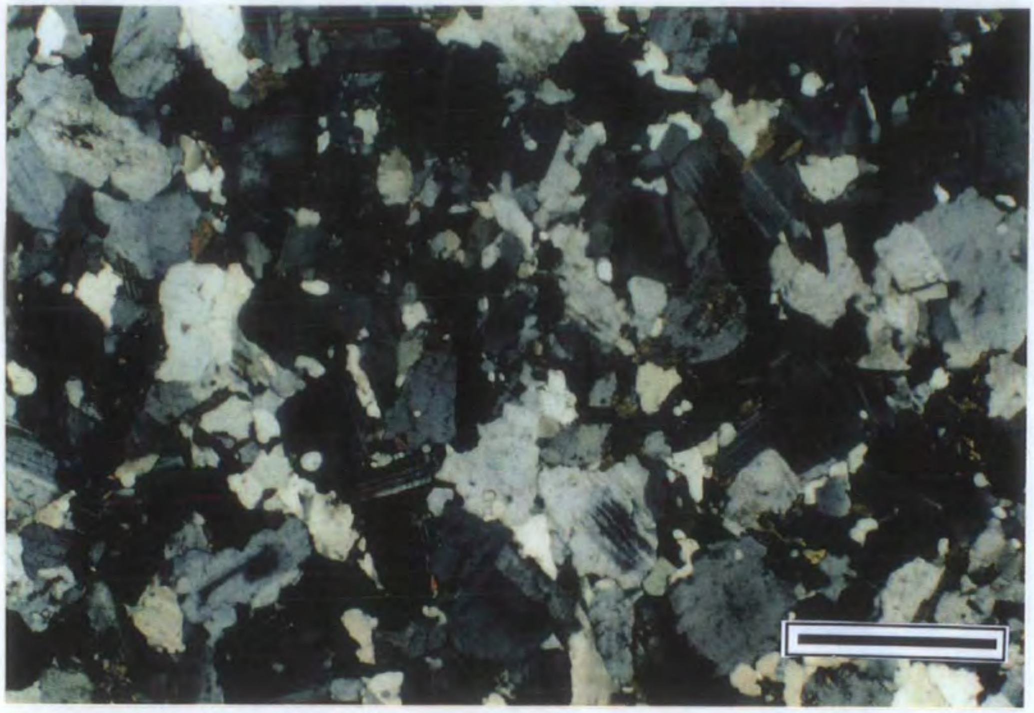


Figure 7:4a:- Photomicrograph of fine-grained granodiorite from the Glendowan Mountains (Map I) showing relatively well developed zoning within the plagioclases (scale bar = 1mm).

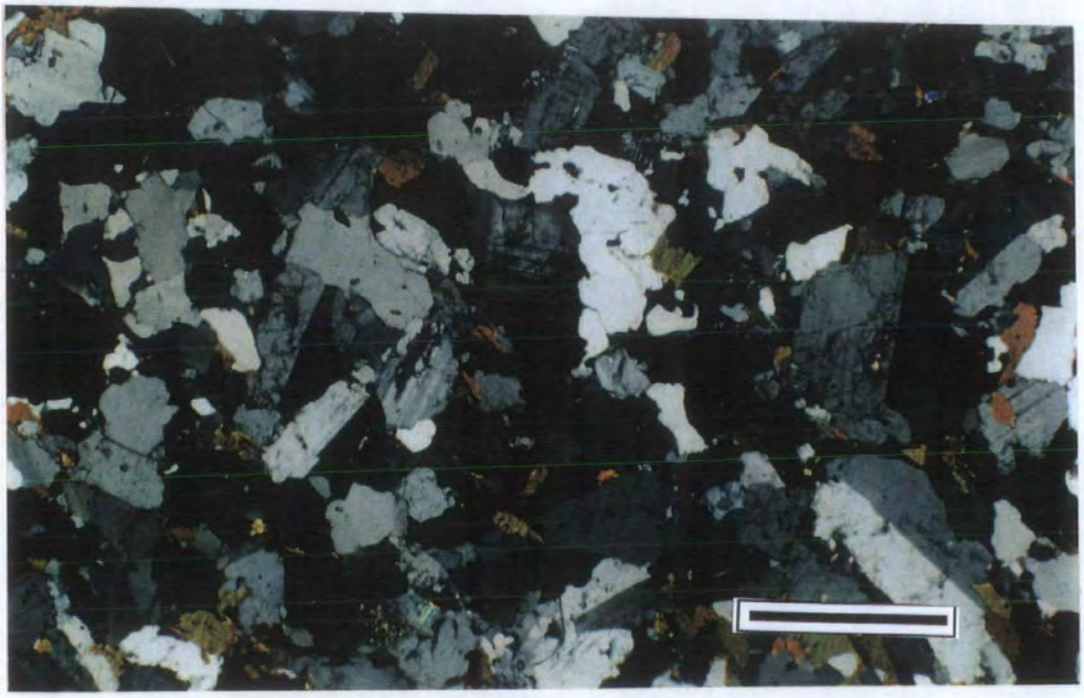


Figure 7:4b:- photomicrograph of granodiorite from the Sruhanavarnis Valley. The proportion of K-feldspar is relatively low with it having an interstitial distribution. Nearby myrmekitic growths indicates its presence (scale bar = 1mm).

It is the authors belief that the latter biotite maybe more analogous to metamorphic biotite, as it clearly cross-cuts the earlier biotite and commonly lies at an inclined orientation to the primary biotite. Due to the higher proportion of K-feldspar in these granodiorites, the degree of chloritisation in biotite is generally much higher than that observed within the tonalites.

Muscovite is present within the granodiorites, tending to occur in association with biotite. The common cross-cutting nature of muscovite flakes implies they may well have crystallised late. In some sections it is also apparent that muscovite has replaced some of the biotite. Only in the heavily sericitised samples are there muscovite flakes within plagioclase feldspar.

The most common accessory mineral within this rock type is epidote which tends to form colourless, equant grains which are mostly seen in close vicinity to, or as inclusions within, biotite. Other accessories include zircon and apatite.

7:3:3 The petrography of the porphyritic monzogranites

The porphyritic monzogranites (plus the aphyric monzogranites of Crobane Hill) comprise the bulk of the granitoid material presently exposed in the central portions of the pluton. As was discussed in Chapter 5, similar monzogranites form more homogeneous masses in the north-western and south-eastern portions of the pluton (especially at the NE end). In all three mapping areas in Chapter 4, several different varieties of monzogranite were encountered which differ in grain size and in the size and colour of the microcline megacrysts. Nevertheless, the mineralogy of the different monzogranites is very similar overall.

Due to the far greater proportion of microcline within these monzogranites the percentage of plagioclase is generally much less, with the average composition for the three mapping areas being: the Sruhanavarnis Valley (An_{22-29}), Glendowan Mountains (An_{21-29}) and Crobane Hill (An_{14-25}). The values for Crobane Hill include the composition of the equigranular CBH 3 monzogranite, which is more felsic than the porphyritic monzogranites and is also believed to be younger, based on field relationships (see section 4:2). The above values are estimates as it was very difficult to find sections in suitable orientation to give accurate extinction angles. Furthermore, within these monzogranites the degree of sericitisation is generally much higher with albite twin lamellae often obscured. The shape of plagioclase is generally subhedral, with the laths broadly rectangular, although the immediate contacts tend to be quite irregular. The inclusions of plagioclase in the larger microcline megacrysts tend to be more euhedral but are also smaller. Around these plagioclase inclusions there are usually thin, more albitic rims which tend to be very fresh in appearance, in contrast with the intensely sericitised mantles.

Berger (1967) also documented these albite rims and found they were only seen when plagioclase was enclosed within K-feldspar. He attributed this feature to exsolution of albitic plagioclase from the K-feldspar lattice, with this exsolved plagioclase accreting to earlier formed plagioclases (Berger 1967 *and references therein*). Within all the porphyritic monzogranites studied, the majority of the plagioclases are unzoned and the rare occasions where zoning is seen it is generally only very weakly developed. In comparison with the tonalite and granodiorites discussed earlier, the degree of alteration in plagioclase is considerably more. The most obvious reason for this appears to be the presence of K-feldspar. A feature which is in accordance with this view is that the plagioclase inclusions in K-feldspar are the most intensely sericitised.

Microcline is the most prominent mineral within the monzogranites and is responsible for giving the porphyritic appearance to these monzogranites. The percentage of microcline within these monzogranites ranges from 20-40%, although the average value is around 31%. Most commonly the K-feldspar forms poikilitic megacrysts which can be up to 12 mm in length, although 6-8 mm is a more typical value. The most common inclusions are plagioclase and biotite, whilst quartz inclusions are generally far less common (see figures 7:5 a & b). K-feldspar occurs as two forms, orthoclase and microcline. Both of these minerals have undergone varying degrees of exsolution, with the development of microperthite. In the orthoclase the perthite is of "flame-like" form whilst in the microcline it tends to be more patchy . In the majority of the porphyritic monzogranites, the K-feldspar megacrysts tend to form quite euhedral crystals which are either square or elongated laths. It must be stated that some K-feldspar does have an irregular crystal form, indicating that it is dominantly interstitial. In orthoclase megacrysts Carlsbad twinning is commonly developed. The twin plane approximately divides the crystal into equal portions and commonly inclusions of plagioclase lie across it. Microcline tends to show the more distinct grid-twinning, although it is not always developed. Myrmekite is extensively developed within the majority of the microcline, where bulbous protrusions of quartz rods (vermicules) in an albitic host extend into microcline. The myrmekitic growths are seen where K-feldspar is in contact with plagioclase, although some contacts are myrmekite free. The origin of myrmekite has been debated for a over a century, but all the details of its formation will not be discussed in excessive detail here. The formation of myrmekite involves the exchange of cations (i.e. K^+ for Na^+ and $Ca2^+$) although deformation processes are believed to enhance its development. Simpson (1985) observed myrmekite development with it tends to occur on K-feldspar margins that are normal to the principal stress direction, due to higher densities of dislocations in these areas. At higher temperatures thermally activated recovery processes take over,

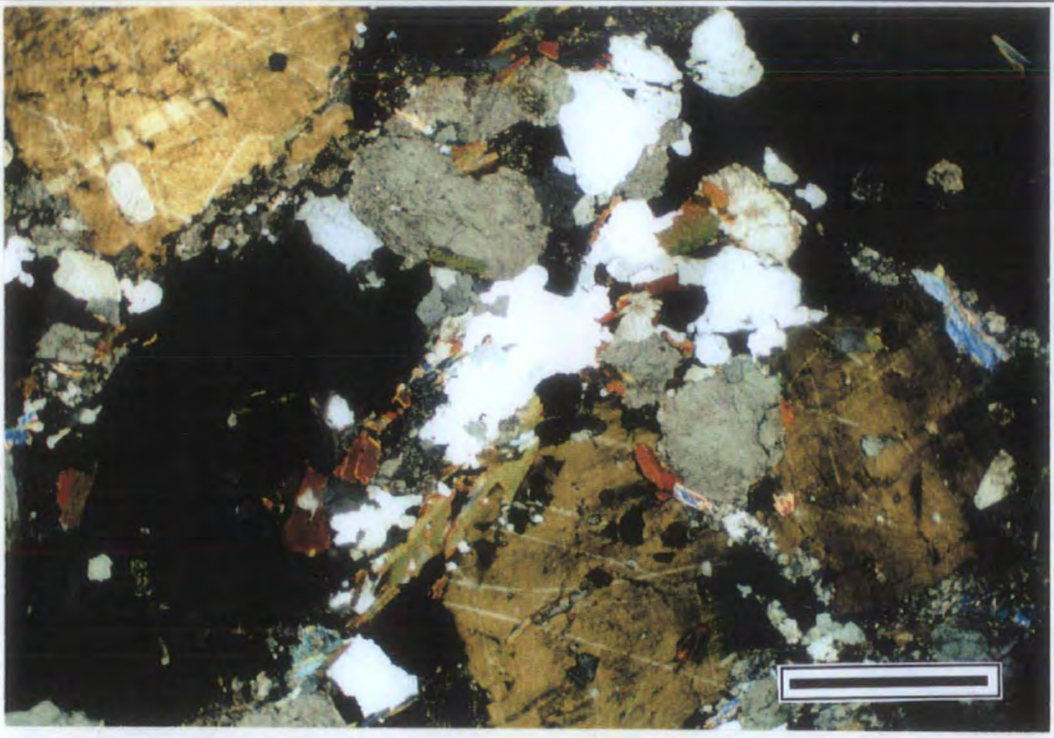


Figure 7:5a:- Photomicrograph of the overall texture of the poprhyritic monzogranites. K-feldspar is yellow (due to sodium cobaltinitrate staining) and commonly forms euhedral (roughly square), poikilitic megacrysts. Contains inclusions of plagioclase, biotite and to a lesser extent quartz (scale bar 2mm)

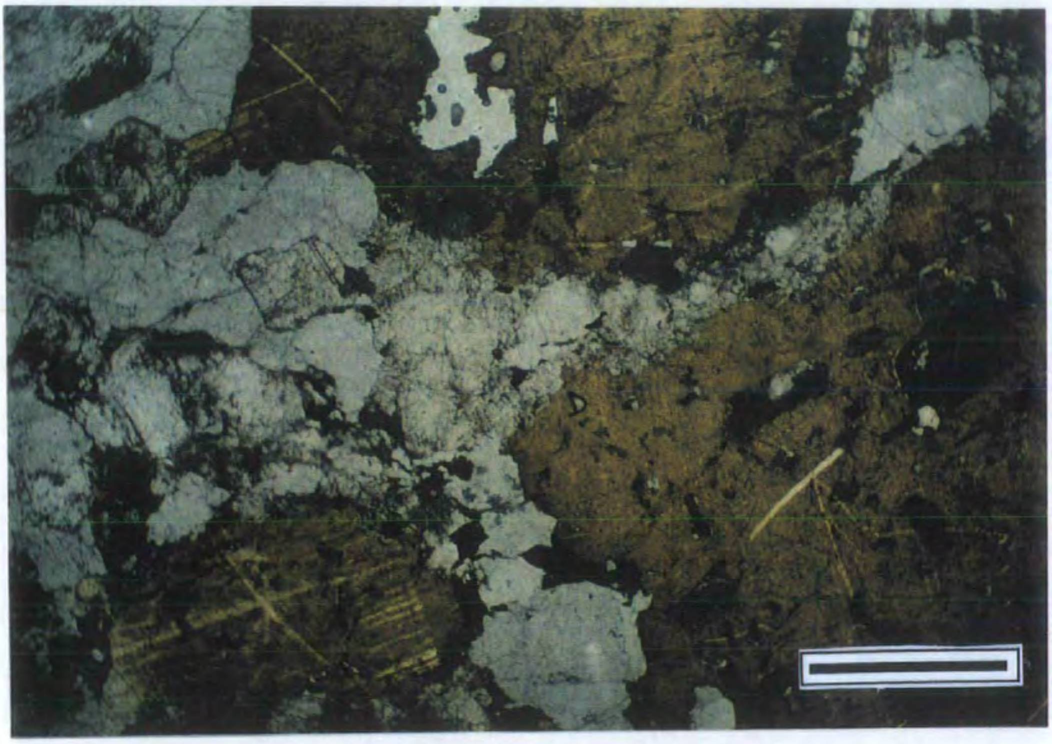


Figure 7:5b:- euhedral plagioclase inclusions within microcline megacrysts (yellow) which are intensely sericitised (darkish mineral), compared to plagioclase outside the microcline, and possess albite rims which tend not to be as sericitised as much (light grey "rind" around plagioclase inclusions) (scale bar = 2mm). (PPL).

with climb and SR recrystallisation becoming more important so that myrmekite tends not to form (Tullis & Yund 1980). Myrmekite development is thus favoured at lower temperatures. With progressive deformation, the myrmekite itself may be further deformed by dynamic recrystallisation, producing fine-grained albite and quartz aggregates which may then be subjected to further grain-sensitive deformation processes such as diffusion creep. The degree of alteration in the K-feldspar is generally quite low, considerably less than that seen within plagioclase feldspar, with cloudy patches of sericite present. From the inclusions within microcline it appears that this mineral started to precipitate during the crystallisation of plagioclase and biotite. The presence of quartz inclusions suggests that this phase may also have been crystallising. Berger (1967) stated that microcline has replaced plagioclase during its growth, with the inclusions being relict remains of this alteration. The euhedral nature of the plagioclase inclusions is a feature that the present author cannot accept as having formed by alteration, as one would expect such inclusions to be far more irregular in shape. The small size of these inclusions is probably due to the microcline engulfing these earlier grains as they grew and hence preventing the plagioclase from further growth whilst, outside the microcline, the plagioclase grew to its present size.

Quartz dominantly forms large, essentially anhedral grains of variable size with the largest crystals being up to 2-3 mm in diameter. Its overall textural distribution implies that the majority of it crystallised later than the microcline and therefore, by inference, the other mineral phases.

Biotite is generally much less common within the porphyritic monzogranites than in the tonalites and granodiorites; on average its modal percentage is 7%, ranging from 3% to 12%. Within all sections it has been extensively altered, with sometimes almost total chloritisation. Within some of the porphyritic monzogranites the red-brown variety is seen e.g. SRU 3: PSBG and the GM 6, although in most of the porphyritic monzogranites the biotite is olive-green in colour.

Muscovite is also quite common, although in most cases it is of secondary origin, deriving from the sericitic alteration of plagioclase with large unaligned plates of muscovite quite common within plagioclase. It is also seen intergrown and commonly cross-cutting biotite.

Accessory minerals are considerably less common within the porphyritic monzogranites with zircons generally less abundant, whilst epidote is rare to absent .

7:3:4 The monzogranites of the homogeneous zones

The following section will address the monzogranites seen in the homogeneous zones discussed in Chapter 4, i.e. the NW pink porphyritic monzogranites and the granitoids which comprise the Crockmore and Binaniller apophyses.

7:3:4:1 NW pink porphyritic monzogranites

In hand specimen the overall texture and appearance of this rock type is very similar to the porphyritic monzogranites seen in the medial regions of the pluton. In thin-section there are also strong similarities with regard to the textural relationships of plagioclase and K-feldspar.

Plagioclase forms euhedral to subhedral crystals ranging from 2-4 mm in length. In some of the euhedral forms there is moderately well developed zoning. In the porphyritic monzogranites of the central areas, zoned plagioclase is a relatively rare occurrence. Inclusions of biotite and quartz within plagioclase are relatively common. The degree of alteration within this mineral is quite low, with sericite tending to occur in the more calcic central regions of the plagioclases.

Microcline shows greater variation in size and form, with the largest crystals up to 11 mm in length. It is these larger grains which tend to be more euhedral-subhedral, whilst the much smaller anhedral microcline appears to be interstitial. This more interstitial microcline also contains very few to no inclusions. Within the larger microclines inclusions of plagioclase (with excellently developed albite rims), biotite and quartz are very common. Due to the higher degree of deformation within these pink porphyritic monzogranites, all of the K-feldspar is in the form of microcline. Furthermore, myrmekite growth is extensively developed within most crystals.

Quartz generally shows much higher degrees of ductile deformation, with moderately strained crystals although around more competent feldspar crystals some of the quartz has recrystallised.

Biotite, which shows varying degrees of chloritisation, has an olive-green colour and defines the foliation within this facies. Muscovite, which is far less abundant, is commonly seen intergrown with biotite and also seen as a sericitic alteration of plagioclase.

The most common accessories are apatite, zircon and allanite. Epidote is relatively uncommon and appears as inclusions in heavily sericitised plagioclase, suggesting it is of secondary origin.

7:3:4:2 The Binaniller apophysis

Where studied in most detail, in the Barnes Gap, the monzogranites which comprise this apophysis are heavily deformed, almost protomylonitic in character. Again the remnant granitic textures within these monzogranites are very similar to the other porphyritic monzogranites which have already been described. The list below describes some of the more important differences (also see figure 5:9).

- Plagioclase tends to be more extensively zoned than in the porphyritic monzogranites of the central regions of the pluton. The more calcic cores tend to be intensely sericitised and contain small flakes of muscovite.
- Microcline megacrysts, which can be up to 5 mm in size, do not have such euhedral forms. Inclusions of plagioclase, biotite and quartz are very common, with the latter mineral being in much greater abundance as inclusions than in the other porphyritic monzogranites. This may suggest that microcline crystallisation has had greater overlap with the crystallisation of quartz. The degree of myrmekitisation is quite extensive within these monzogranites.
- Biotite has an olive-green to khaki-brown colour and has been extensively altered and deformed, and in some parts of the section it has a "shredded" appearance. Muscovite appears to have mainly derived from alteration of plagioclase and biotite and is commonly seen along the foliation planes, most notably the C' (S_8) planes.
- One of the more important differences between the monzogranites of the Binaniller apophysis and the porphyritic monzogranites of the central and north-western regions is the greater abundance of epidote in the former. It is most commonly seen as large crystals, up to 1 mm in size, in the close vicinity of biotite although to a lesser extent it is seen in the centre of plagioclase crystals. In some of the larger epidotes the core of the crystal is composed of darker allanite.

  Other accessories include zircon and apatite.

7:3:4:3 The Crockmore apophysis

In hand specimen this monzogranite has a dominantly equigranular appearance and is often intensely deformed. The strain generally decreases away from the Crockmore Septum (NW boundary) and and away from the SE contact of the pluton, allowing the petrography of relatively unstrained granitoid to be studied (see fig. 5:7).

Plagioclase forms euhedral to anhedral laths up to 2-3 mm in length. The composition of plagioclase ranges between An_{21-27}, with this value taken from zoned crystals. Zoning within plagioclases is very well developed within most of the crystals, most notably the euhedral inclusions within microcline megacrysts.

Microcline tends to form very large subhedral crystals which can be up to 15 mm in size. Within these megacrysts are large inclusions, up to 2 mm in diameter, of biotite and plagioclase, with quartz tending to be a relatively uncommon inclusion. The larger size of the inclusions within the poikilitic megacrysts is the probable reason for this granitoid appearing equigranular at the outcrop scale. Furthermore the white colour of both the feldspars would also disguise its porphyritic nature.

Within all the granitoids of the Crockmore apophysis the biotites have a red-brown colour, indicating they are relatively titanium-rich. The degree of chloritisation is only very minor within this apophysis. Within the undeformed Crockmore Granite muscovite is quite scarce (apart from flakes in the cores of plagioclase crystals). In the more deformed parts of the apophysis, where muscovite is more abundant, it appears to have developed from the alteration of biotite and plagioclase.

Accessory minerals are generally rare, with zircon being the most abundant. Within the Main Donegal Granite zircon tends to be more abundant in the granitoids which have the red-brown coloured biotites, a feature Berger (1967) also commented on. Apatite is present to a lesser extent, occurring as pseudohexagonal inclusions within biotite. Epidote was not observed in any of the sections studied from the Crockmore apophysis.

7:3:5 The petrography of the granitoids at Galwollie Hill and the Trawenagh Bay Pluton

The granitoids of the Galwollie Hill area are situated close to the north-western boundary of the Main Donegal Granite Shear Zone and as a result are intensely deformed. It is only in the earlier granitic phases that the deformation is less intense and the original petrography is preserved petrographic. The GWH 1 tonalite, which is relatively undeformed, has been addressed in the earlier sections concerning the tonalites within the pluton. The porphyritic GWH 2 monzogranite is also texturally very similar to the porphyritic monzogranites already described.

In chapter 6 it was stated that the GWH 3 and GWH 4 & 5 monzogranites may belong to the G I and G II pulses, respectively, of the Trawenagh Bay Granite. The GWH 3 monzogranite is texturally quite different to the MDG porphyritic monzogranites, due to the presence of more abundant, smaller K-feldspar megacrysts and the overall coarser grainsize of the groundmass. Therefore in hand specimen these granitoids are weakly porphyritic. The GWH 4 and GWH 5 monzogranites are characterised by their high proportion of muscovite which, from textural relationships, appears to be of primary igneous origin. In Map C this G II phase (GWH 4 & 5) was intensely deformed, appearing almost protomylonitic in character with igneous fabrics having been obliterated. For this reason the petrography of the G I and G II phases will be described from areas as far away as possible from this zone of highly strained and reddened granite.

G I:- in hand specimen this monzogranite has a weakly porphyritic appearance, due to the presence of poikilitic microclines which can be up to 9-10 mm in diameter. Common inclusions are biotite, quartz and plagioclase. In all samples the K-feldspar is in the form of microcline. Myrmekitic growths are common within the majority of the microcline crystals. Plagioclase is on the whole quite euhedral, forming laths up to 3-4 mm in length, and shows weak to moderately developed extinction. The composition of plagioclase ranged from An_{18-24}, with some crystals commonly showing well developed zoning. In comparison with the porphyritic monzogranites of the MDG, which are petrographically very similar to G I, it was observed that plagioclase zoning was weak to absent. Sericitic alteration is quite extensive within plagioclase, with small flakes of muscovite commonly occupying the central regions of these crystals. Biotite (olive-green variety) has been extensively chloritised, especially along the (001) basal cleavage. Muscovite is quite common, appearing as small flakes which are frequently seen intergrown with biotite or otherwise occurs within plagioclase crystals. Within this facies the muscovite appears to have been derived by alteration of plagioclase and biotite. Epidote tends to be quite a common accessory within all of the samples of the G I monzogranite and is usually seen in association with biotite. Other accessories include apatite and zircon.

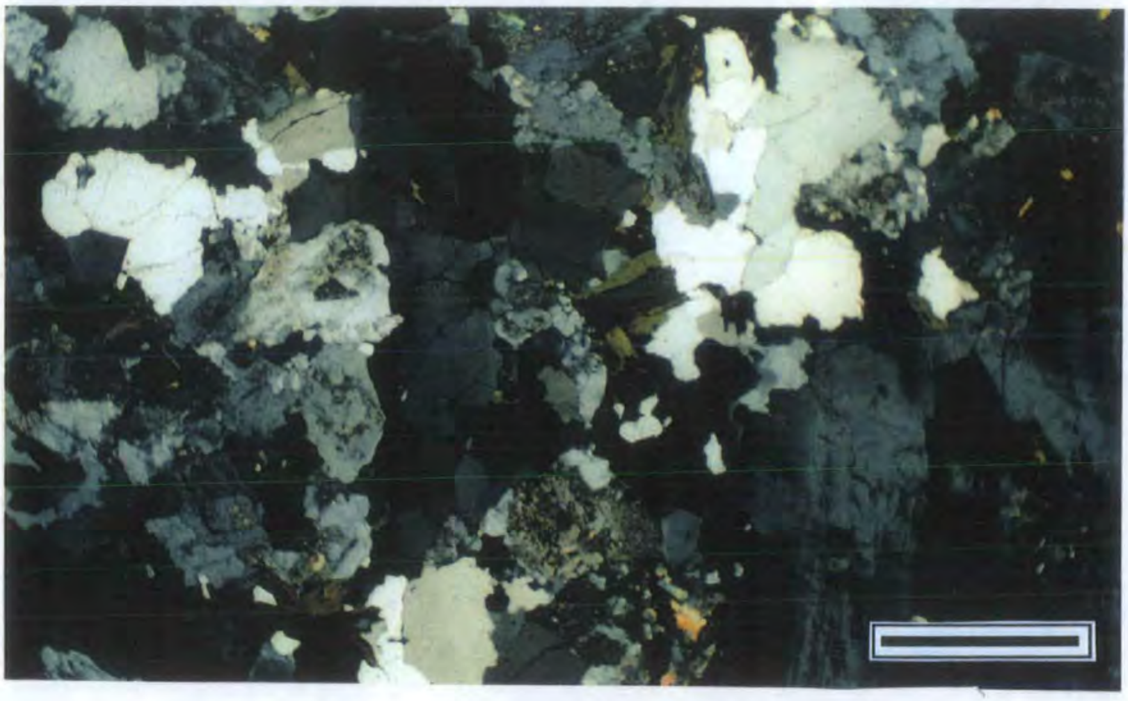


Figure 7:6:- photomicrograph of very weakly deformed G I with quartz appearing unstrained. Microclines are quite large but commonly contain inclusions which results in this monzogranite appearing equigranular at the outcrop scale (scale bar = 2mm).

G II:- Plagioclase tends to form relatively euhedral grains, up to 3 mm in length, and often shows weak to moderate zoning. K-feldspar forms crystals up to 7 mm in size and occurs either as microcline or microperthitic orthoclase with the latter polymorph commonly showing Carlsbad twinning. The orthoclase (which is variably converted to microcline) tends to be the most euhedral and this may suggest it was an early crystallising phase. In contrast, microcline is more irregular shaped and appears to be dominantly interstitial. Inclusions of biotite and plagioclase are quite common whilst quartz inclusions are generally more rare. Olive-green biotites tend to be intensely altered, almost totally chloritised in some samples; a feature which is partly attributable to the intense deformation in the Galwollie Hill area and to the relatively high proportion of K-feldspar. Muscovite occurs as large flakes up to 2 mm in size, although it also forms large clusters and its textural distribution implies that it has crystallised from a granitic magma. In the highly deformed GWH 5 monzogranite the smaller muscovite are commonly seen along the S_8 shear planes and in these sections it appears to have derived from the alteration of biotite. Accessory minerals are relatively uncommon in the GWH 4 and GWH 5, apart from occasional epidotes which are always seen in close association with biotite.

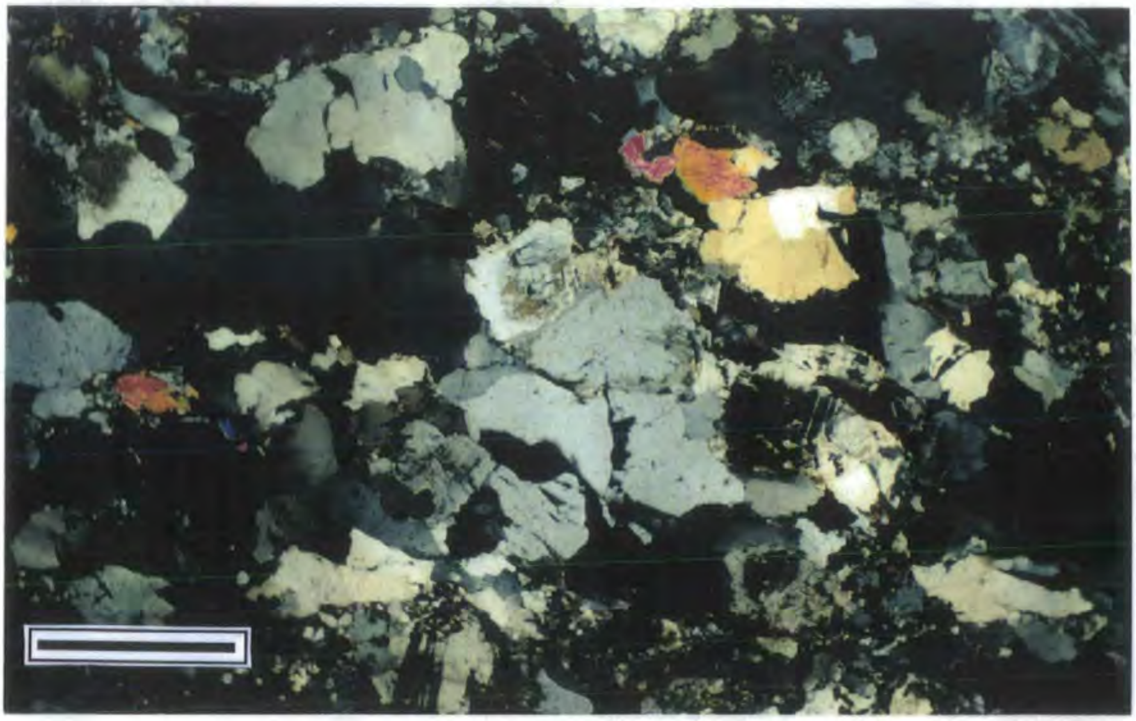


Figure 7:7:- photomicrograph of moderately deformed G II monzogranite, although primary magmatic fabrics are strongly overprinted. Muscovite tends to occur as relatively large plates and appears to have crystallised from a melt and not formed by alteration of other mineral phases (scale bar = 1mm).

7:4 Overall geochemical variation within the Main Donegal Granite

As well as modal classification, the granitoids of the Main Donegal and Trawenagh Bay plutons have also been classified using several geochemical methods which use varying combinations of major oxide proportions within analysed granites. At this point in this chapter it is important to state that the names given to the granitoids are based on modal analyses i.e. the proportions of plagioclase feldspar, alkali feldspar and quartz. These names will be maintained when geochemical variation within the granitoids is discussed in subsequent sections, although any cryptic variation will refer to the geochemical data. Three different classification systems using the geochemical data have been used:

(1) the "K_2O v silica" diagram of Rickwood (1989) and Le Maitre et al. (1989)

Within this plot (figure 7:7) the three main granitoid types within the Main Donegal pluton are differentiated from one another. The porphyritic monzogranites lie mainly within the "high-K, calc-alkaline" series whilst the tonalites, and to a lesser extent, the granodiorites lie within the "calc-alkaline" series. The granodiorites show a greater spread, with this feature controlled mainly by the presence of K-feldspar and also biotite. The presence of some granodiorites in the "tonalite" field is believed to result from erroneous identification in the field, i.e. there may have been some finer grained tonalites which resemble the granodiorites in appearance.

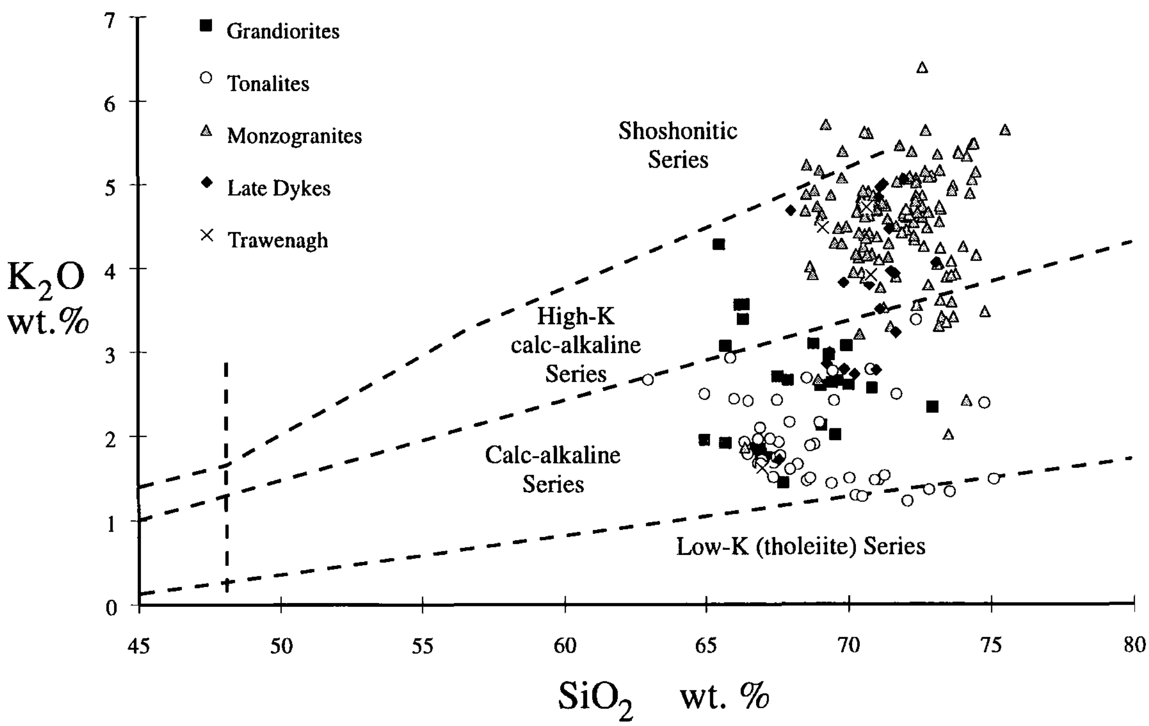


Figure 7:7:- K_2O versus Silica for the granitoids of MDG and TBG plutons (modified from Le Maitre *et al.* (1989) and Rickwood (1989).

(2) the "TAS" (Total Alkalis versus Silica) of Wilson (1989)

The TAS plot of Cox *et al.* (1979) was originally devised for the sub-division of volcanics into alkaline and tholeiitic series. Wilson (1989) modified this diagram for use in classifying plutonic rocks. Rollinson (1993) states that the TAS plot of Wilson (1989) uses the boundaries defined by Cox *et al.* 1979 which was intended for volcanic rocks and therefore the boundaries are not quite accurate for plutonic equivalents. The TAS plot for the granitoids in this study plots the majority of the tonalites and granodiorites in the "granite" and "quartz diorite" fields whilst the majority of the monzogranites plot in the "alkali granite" field. The heavy curved line in figure 7:8 separates subalkalic rocks from alkalic rocks. Therefore in this classification the majority of the monzogranites are alkalic in nature whilst all of the tonalities and granodiorites are subalkalic.

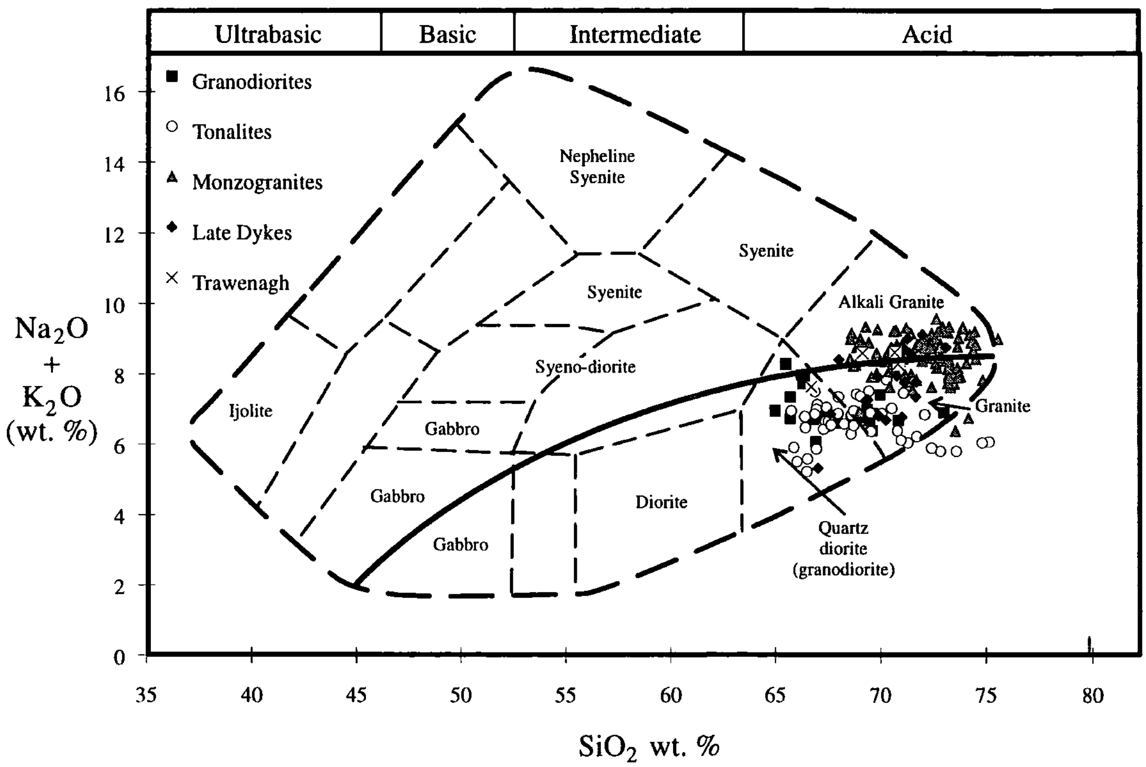


Figure 7:8:- Total Alkalis v Silica (TAS) for the granitoids of the MDG and TBG plutons (*after* Wilson 1989).

(3) the "*R1-R2*" plot of *de la Roche et al.* (1980)

The R1-R2 plot devised by De la Roche *et al.* (1980) is more representative than the TAS plot because it incorporates the majority of the major cations found in igneous rocks (see figure 7:9). Within this plot the tonalites and granodiorites lie mainly within the "granodiorite" field and generally they are intermingled with one another. The monzogranites form a separate group which mostly lie within the "monzogranite" field and to a lesser extent within the "syenogranite" and "granodiorite" fields. In summary the R1-R2 plots clearly differentiate the early tonalites and granodiorites from the porphyritic monzogranites.

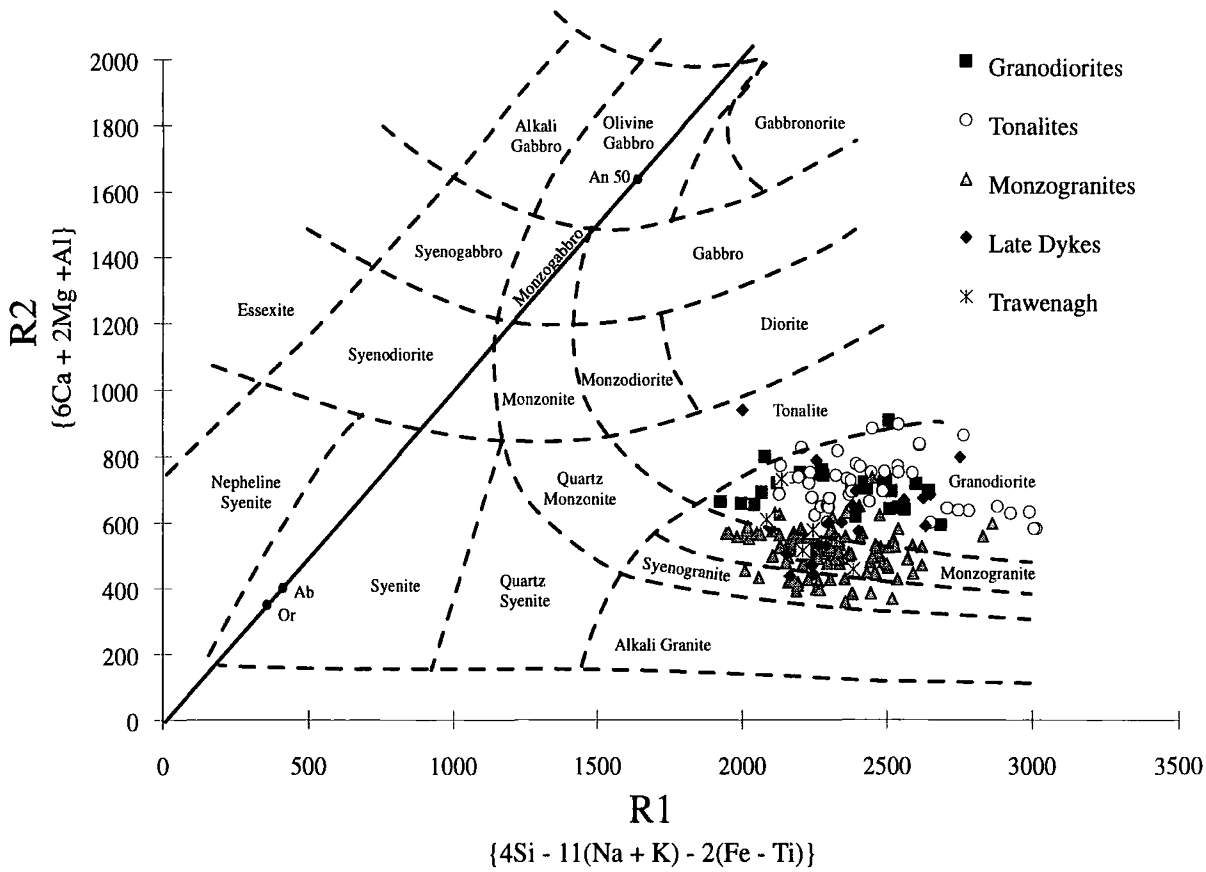


Figure 7:9:- R1-R2 plot of the granitoids within the MDG and TBG plutons based on millicationic proportions (after De la Roche *et al.* 1980).

Therefore the majority of the granitoids within the pluton show calc-alkaline affinities. Using the geochemical data, the granitoids were also tested for peraluminous ((K+Na+2Ca)/Al <1) and peralkaline ((K+Na)/Al >1) affinities. None of these granitoids were peralkaline, a feature confirmed by the petrography, although some granites were peraluminous. Thin-sections of these peraluminous monzogranites confirm the geochemical data, with muscovite usually in quite high abundance (~4% of the mode).

7:4:1 Major and trace element variation

The major element chemistry of the granitoids which comprise the Main Donegal and Trawenagh Bay Granites is shown in the Harker plots, where SiO_2 is plotted against all the other major oxides (figure 7:10). Furthermore, in figure 7:11 the granitoids have been plotted against a series of trace elements in an attempt to identify more subtle variations. The overall trend for all the granitoids sampled is the older granodiorites being the most "primitive"*[1], with the tonalites and monzogranites becoming progressively more "evolved" ‡[2] with time. All of the major oxides, apart from K_2O, decrease as the percentage of SiO_2 increases. The increased K_2O in the later porphyritic monzogranites relates to the presence of large K-feldspar megacrysts. Amongst this overall linear trend there are more subtle trends developed within each of the three main granitoid phases. Within the granodiorite field the silica content ranges from 65-73%, with major oxides tending to show a decrease with increasing silica. This similar pattern is seen within the tonalites (SiO_2 = 66-75%) and within the monzogranites (68.5-75.5%). This trend implies that some degree of fractionation may have occurred within each of the three main granitoid components of the Main Donegal Granite. The following section discusses the geochemical variation within some of the individual mapping areas and may give a more spatial distribution to the internal variation seen within the three major granitoid phases that comprise the bulk of the pluton.

The small intrusive dykes, which are mostly later than the major granitoids phases, display a more random cluster although the majority of them are geochemically similar to the monzogranites and, to a lesser extent, the tonalites The samples obtained from the Trawenagh Bay pluton are virtually identical in composition to the monzogranites of the Main Donegal pluton although the G II member does show less similarities. As was mentioned in section 7:3 the granodioritic sample from the Trawenagh Bay pluton is believed to be a xenolith belonging to the Main Donegal pluton.

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* In this thesis the term "primitive" refers to granitoids which are characterised by minerals which generally have higher liquidus and solidus temperatures. Such minerals include calcic plagioclase, biotite and numerous accessories. Geochemically these "primitive" granitoids will have lower silica contents whilst other major oxide values will be relatively high. During fractional crystallisation one would expect the more "primitive" granitoids to crystallise first as the solidus temperatures for its constituent minerals are the highest. The reverse situation is true for partial melting.

‡ The term "evolved" in this thesis will refer to granitoids typified by minerals which have low liquidus and solidus temperatures, i.e. quartz and alkali feldspar. These granitoids are therefore characterised by higher silica levels and tend to be more depleted in the other major oxides, possibly except for K_2O and Na_2O. During fractional crystallisation "evolved" granitoids may be the last to form as the constituent minerals will only crystallise when temperatures are sufficiently low enough. During partial melting more evolved melt compositions will form first.

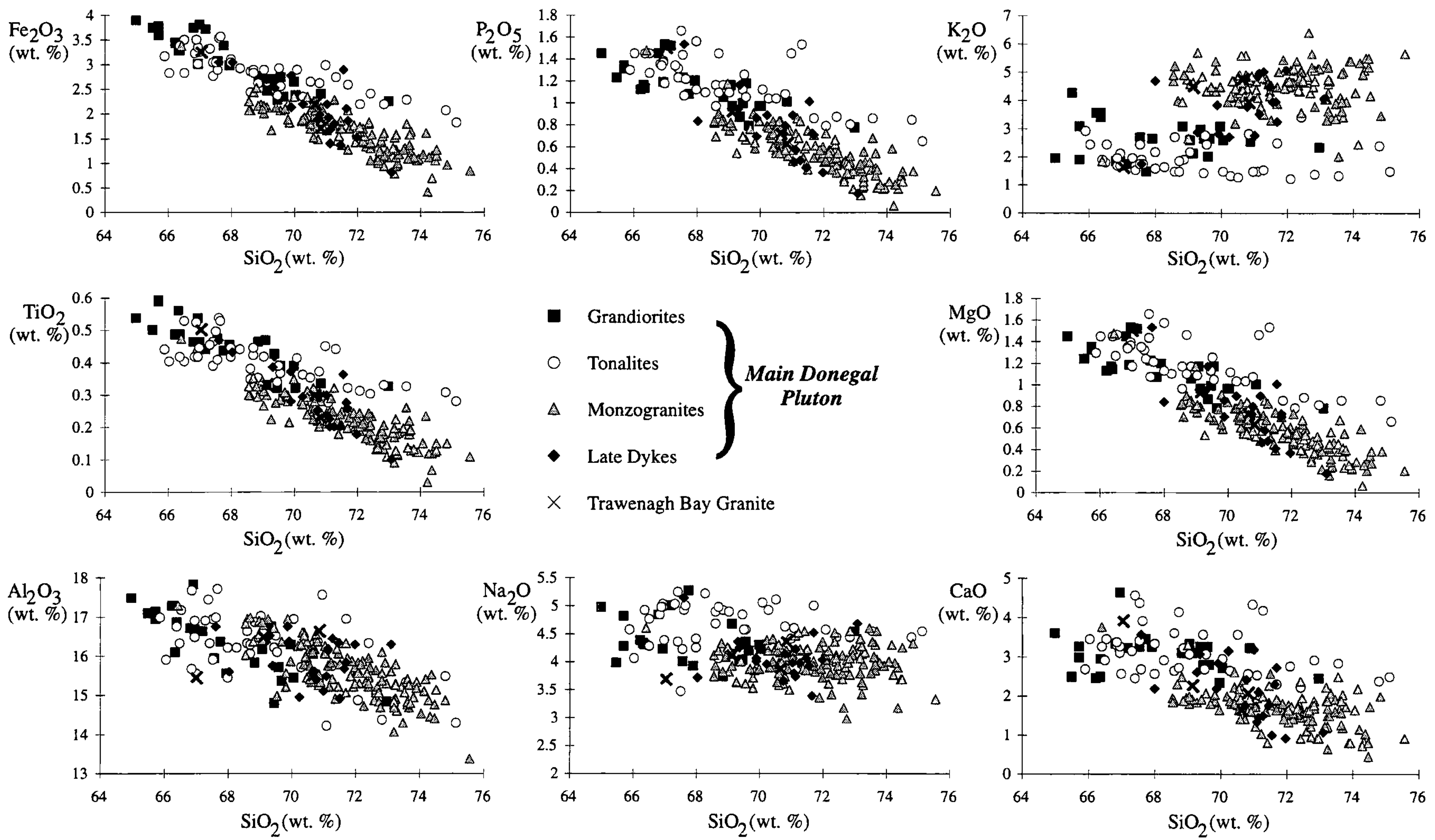


Figure 7:10:- Harker plots for the granitoids which comprise the Main Donegal Granite

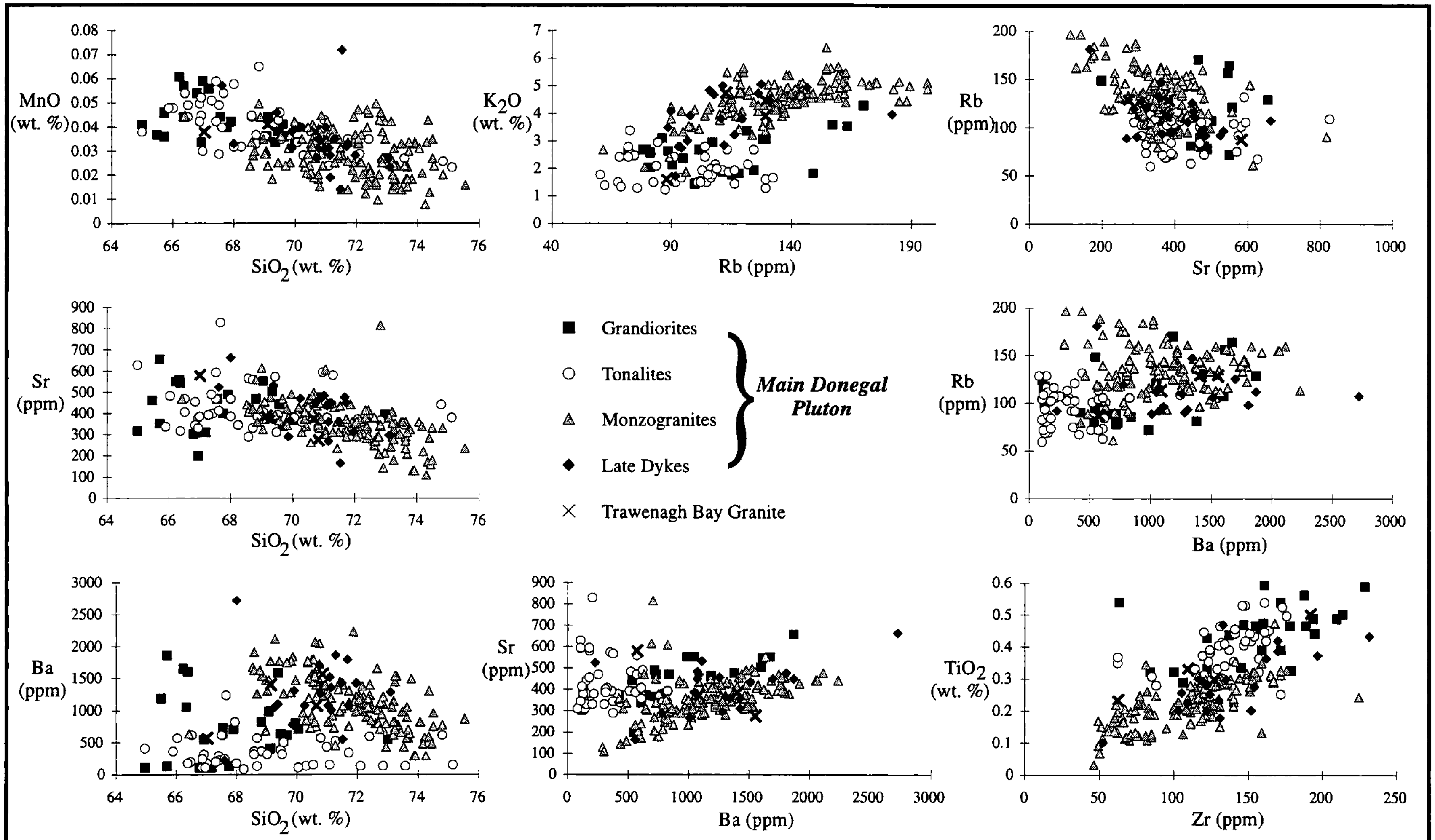


Figure 7:11:- Various Harker and trace element plots for the granitoids of the Main Donegal Granite

7:4:2 Geochemical Variation in the heterogeneous central zones

In Chapter 4 it was noted that in all three areas (i.e. Crobane Hill, Sruhanavarnis valley and Glendowan Mountains) there was an overall similar style of emplacement. The oldest granitoids tending to be granodiorites and tonalites (which in themselves may belong to two separate phases of intrusion), which are commonly found as rafts within later and generally more abundant porphyritic monzogranites. In all three areas, most notably the Sruhanavarnis Valley and Glendowan Mountains, considerable internal variation is seen within the porphyritic granites. This section will initially address the geochemistry of the individual areas, whilst the latter part of the section will summarise all three areas and whether or not the geochemistry supports the field evidence and also to see if it identifies any more subtle, underlying cryptic variation.

7:4:2:1 The Sruhanavarnis Valley

The geochemical samples collected in the Sruhanavarnis valley mainly come from the Map B area where the age relationships and spatial distribution of the granitoids are well known (see appendix B for exact locations). The majority of the samples are from fresh hand specimens, although at the more north-western end of the mapping area many of the samples had to be collected by drilling a series of cores at each sample point. In figure 7:12 the major element geochemistry has been plotted in a series of Harker plots, with SiO_2 as the abscissa, in order to show the main geochemical characteristics of the granitoids in this area. These plots indicate that the oldest granodiorites are overall the most primitive with the younger porphyritic granites becoming progressively more evolved. Despite this general trend there are a number of interesting features:-

1) Between SRU 1 and SRU 2 there is some degree of overlap, although the majority of the former are slightly more primitive. The overlap may be due some of the tonalites being finer grained and resembling SRU 1 at the outcrop level and vice versa where some of the SRU 1 granodiorites were coarser and hence called SRU 2.

2) The SRU 2 granites show a wide variation in silica content, ranging from ~65 to 74% and almost encompasses the entire range seen in all the Sruhanavarnis Valley granitoids. The more silica-rich SRU 2 samples can be distinguished in thin-section from the more intermediate SRU 2 members by the colour of the biotite crystals. As in the more silica-rich samples, this mineral has a red-brown colour whilst the more intermediate tonalites have olive-green biotites. As already mentioned the red-brown colour is due to higher relative amount of Ti^{4+} in respect to Fe^{3+}. In figure 7:12 the plot of SiO_2 *v* TiO_2 shows this higher amount of Ti^{4+} in the SRU 2 tonalite, relative to other monzogranites with similar SiO_2 contents. These more silica-rich tonalites

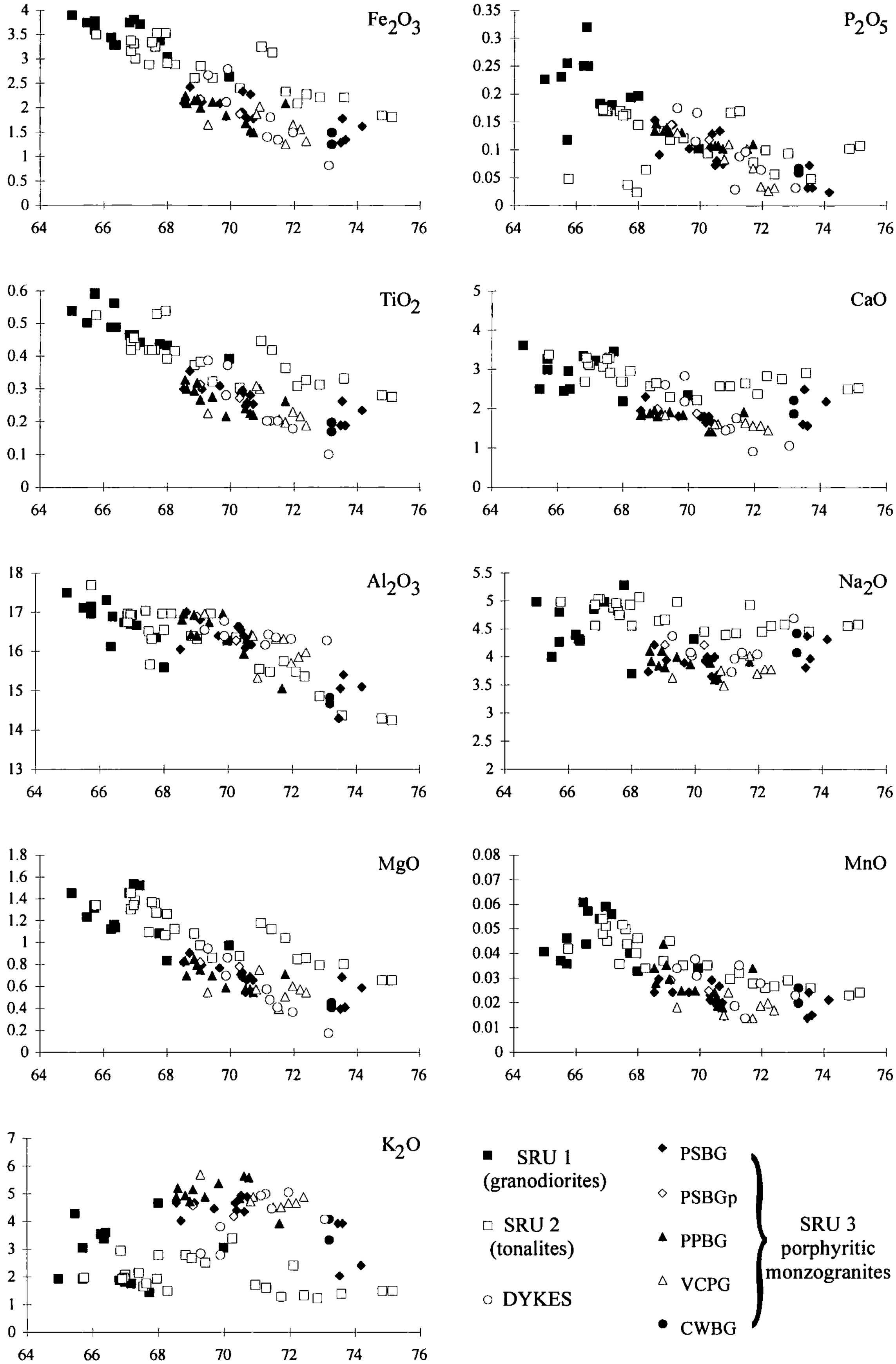


Figure 7:12:- Harker plots for the granitoids of the Sruhanavarnis Valley (wt. %). (Horizontal axes (x) is SiO_2)

(based on geochemistry) also generally have much higher proportions of MgO, CaO and Na_2O, in comparison with the porphyritic monzogranites with similar silica content.

The chemical variation within the SRU 2 tonalites may be a feature either inherited prior to emplacement (at the source or during ascent) or it may have occurred at the level of emplacement (i.e. contamination). The location of these granitoids within a major raft-zone (Glenveagh 3A), where pelite forms a major component, makes the latter option a possibility. To investigate possible contamination a couple of pelites (taken from squares L1 and Q2 of Map B) were analysed for trace elements and were plotted against all the major granitoid phases on a spider diagram (figure 7:13a) to try identify any pelitic involvement.

One would expect the earlier granitoid phases, especially SRU 1 granodiorite and to a lesser extent the SRU 2 tonalite, to have had more interaction with the country rock, as it was these granitoids which first intruded the country rocks of this area. The later porphyritic monzogranites would possibly be less contaminated as they have been dominantly emplaced into earlier formed granodiorites and tonalites and have therefore had less contact with the pelite rafts at the present level of exposure. This theme was investigated further, by plotting sets of samples of the SRU 1 granodiorite (figure 7:13) and SRU 2 tonalite (figure 7:13c) taken at varying distances from the rafts. In theory one would expect the SRU 1 and SRU 2 samples closest to the rafts to be the most contaminated but the plots do not show any clear evidence to suggest this. Therefore it is clear that any Dalradian input to the Main Donegal Granite magmas must have taken place at a deeper level than the present exposure.

The porphyritic monzogranites generally show minor variations in their major element chemistry, tending to group together on the Harker plots with slight overlap between the SRU 3 variants. On the whole the PSBG, PSBGp and PPBG tend to be very slightly more primitive than the VCPG and CWBG monzogranites.

The trace element data were plotted on spider diagrams and have been normalised to primordial mantle (values from McDonough *et al.* (1992)) and also to the most "primitive" granitoid seen within the Sruhanavarnis Valley (figure 7:14 a & b). In the primordial mantle normalised plot the overall trend within all of the granitoids shows that the HFSE become progressively more depleted. A number of anomalies are present:- i) all of the granitoids show a negative Nb anomaly, a feature generally typical of most continental crust rocks (Tarney & Jones 1994). ii) The tonalites show a marked negative anomaly in the amount of Ba.

In the trace element plot where the granitoids are normalised to the SRU 1 granodiorite the later more-evolved monzogranites tend to be depleted in the HFSE.

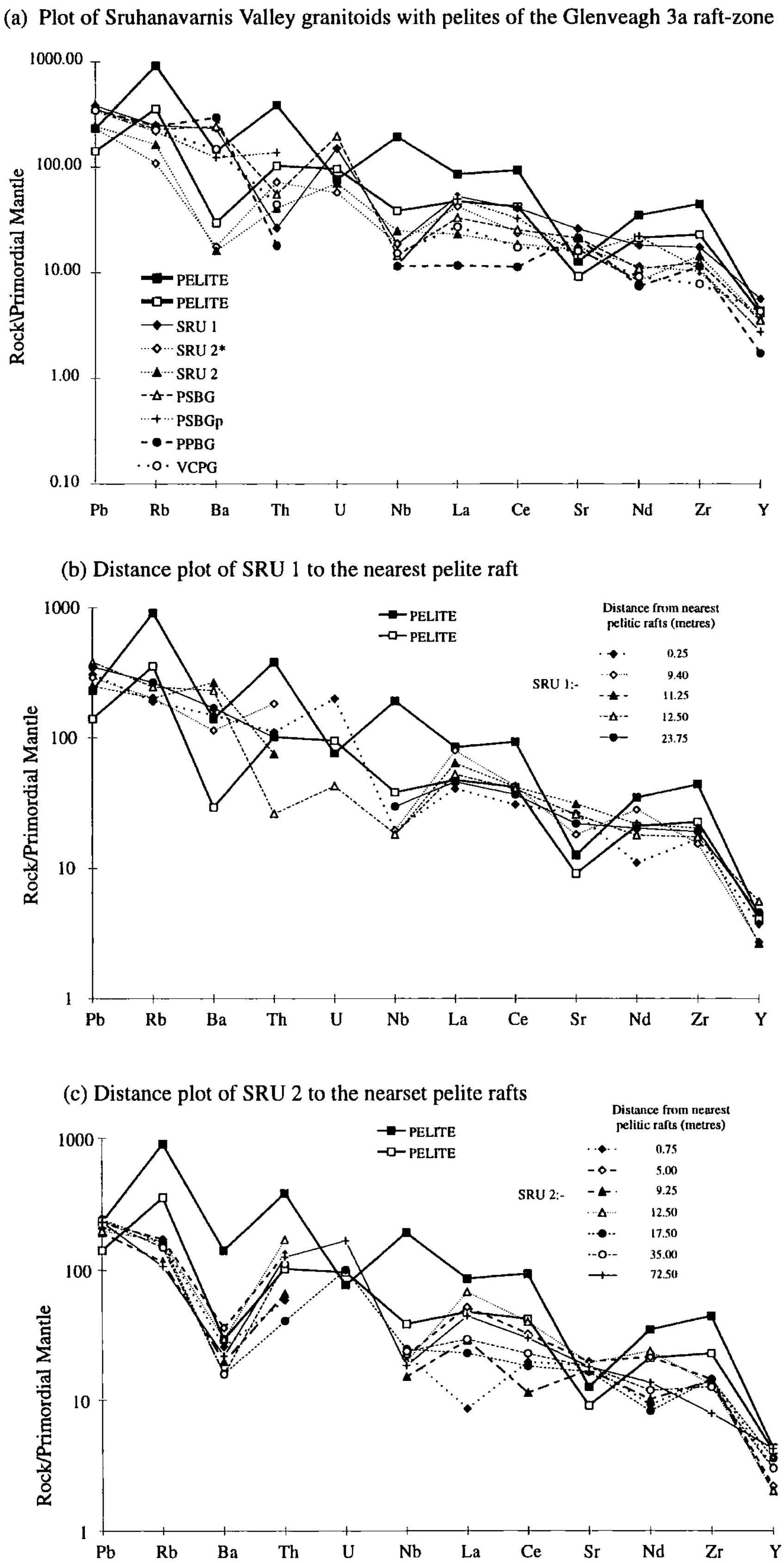
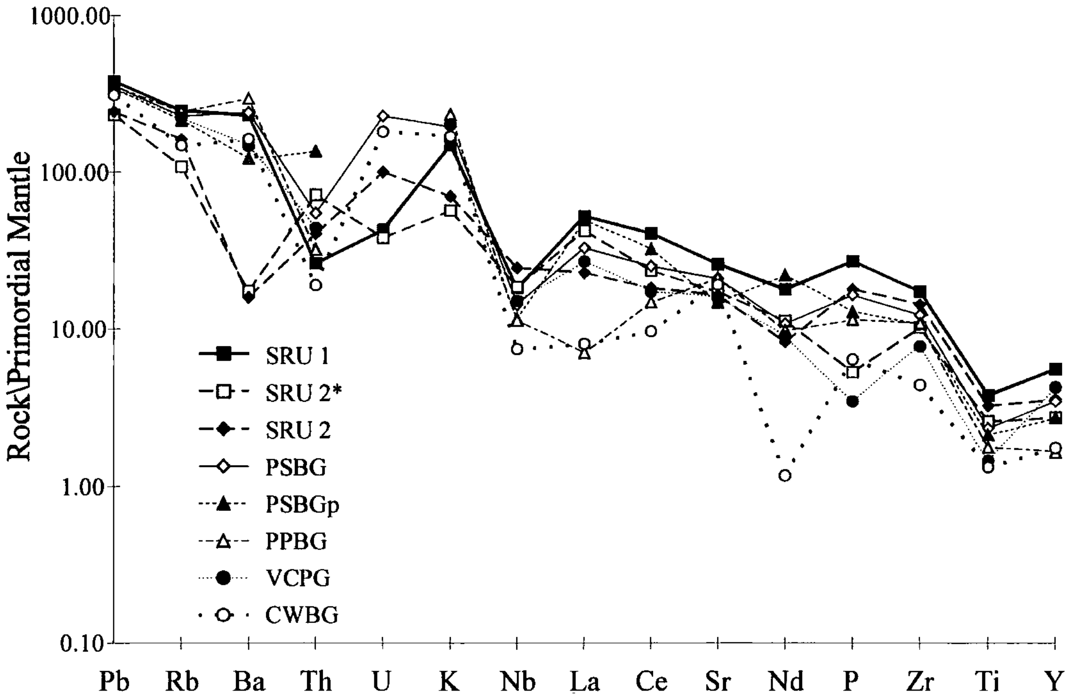
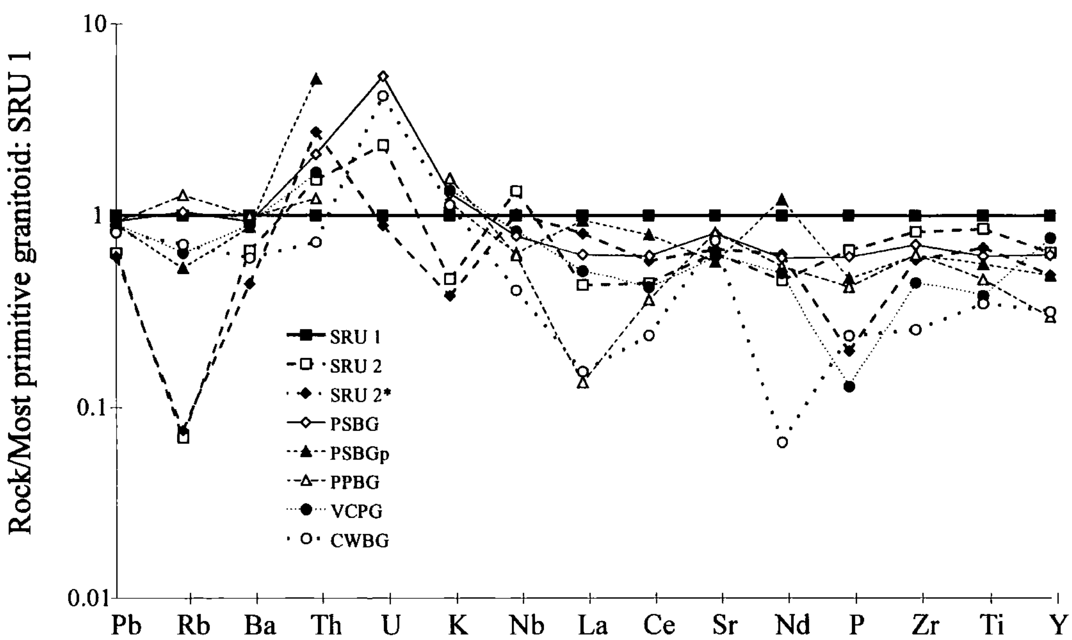


Figure 7:13:- Spiderdiagrams investigating the possibility of contamination within the granitoids of the Sruhanavarnis Valley. (Normalised to primordial mantle McDonough et *al.* 1992)



(A) Trace element variation within the "SRU" granitoids normalised to primordial mantle (after McDonough *et al.* (1992). P^{5+} from Sun (1980)).



(B) Trace element variation within the "SRU" granitoids normalised to most prmitive granitoid (SRU 1).

Figure 7:14:- Spider diagram plots showing trace element variation within the Sruhanavarnis Valley granitoids.

Partition coefficients for these elements imply they mainly go into accessory phases, such as zircon, allanite, apatite and also the major phase biotite. This feature is in accordance with the petrography of these monzogranites, with accessory minerals becoming far less abundant within the younger porphyritic monzogranites. In comparison with the SRU 1 granodiorite the tonalites display a marked negative Rb anomaly. In both plots the SRU 3: CWBG monzogranite is strongly depleted in the more HFSE implying fractionation of accessory minerals, which commonly accept these elements, may have already occurred prior to the formation of this monzogranite, i.e. the SRU 3: CWBG is more evolved than the other monzogranites.

7:4:2:2 Crobane Hill

On Crobane Hill fewer geochemical samples were collected as the granitoids were generally much easier to differentiate between at the outcrop scale. Figure 7:15 are Harker plots for the major element data with SiO_2 again as the abscissa. The plots show a relatively simple trend, with the majority of the younger granitoids becoming progressively more evolved. The exception to this is the CBH 4 disrupted dyke which is the most primitive granitoid in this area. The CBH 4* tonalite which is identical, in regards to petrography and appearance, to the CBH 4 tonalite also has almost identical major element chemistry. The majority of the plots form a linear trend, apart from Al_2O_3. This scatter may well be due to analytical error, as this oxide was quite variable due to analysis of powdered pellets rather than fusion discs. Otherwise all oxides, except K_2O, decrease as the percentage of SiO_2 increases.

Similar trace element plots to those in the Sruhanavarnis Valley were also done for the granitoids of Crobane Hill. The primordial mantle normalised plots (figure 7:16a) are very similar to that seen within the Sruhanavarnis Valley although there are subtle differences. The CBH 4*\CBH 4 tonalites show less of a Ba negative anomaly than the tonalites of the Sruhanavarnis. The CBH 3 monzogranite, which is the youngest major phase in this area, is more strongly depleted in the HFSE, a feature which is more apparent in the plots where the granitoids have been normalised to the most primitive CBH 4 tonalite (figure 7:16b). In this latter plot all of the younger monzogranites show strong positive anomalies of Rb, Th and K.

7:4:2:3 The Glendowan Mountains

Geochemical samples in the Glendowan Mountains were collected over a greater area than the samples of the Sruhanavarnis Valley and Crobane Hill. The majority of the samples were collected from Maps I and II, although samples were collected from the summit area of Croaghacullin and the SW flanks of Moylenanav. Two samples were collected from the summit region of Leahanmore (LEAH),

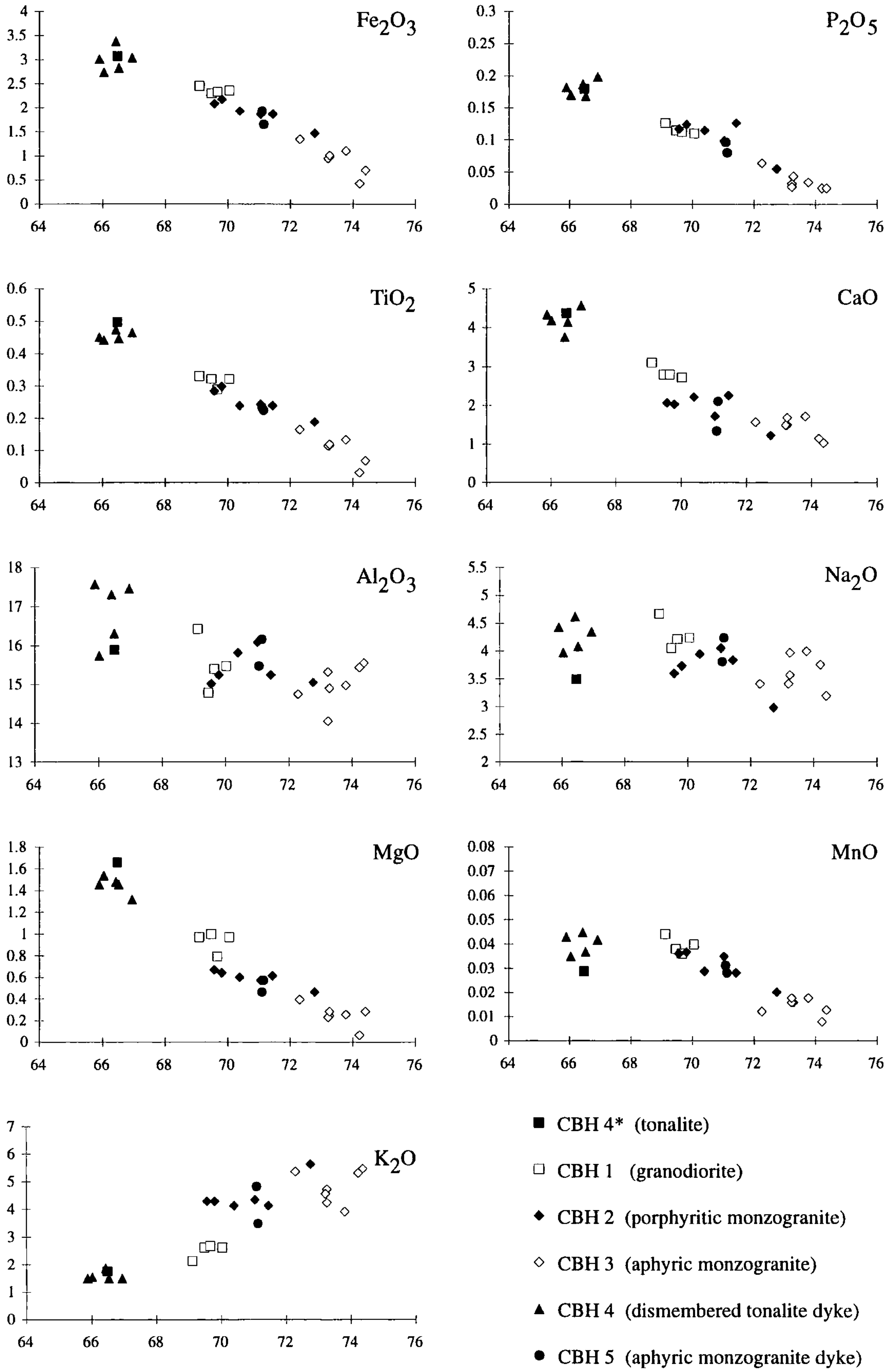
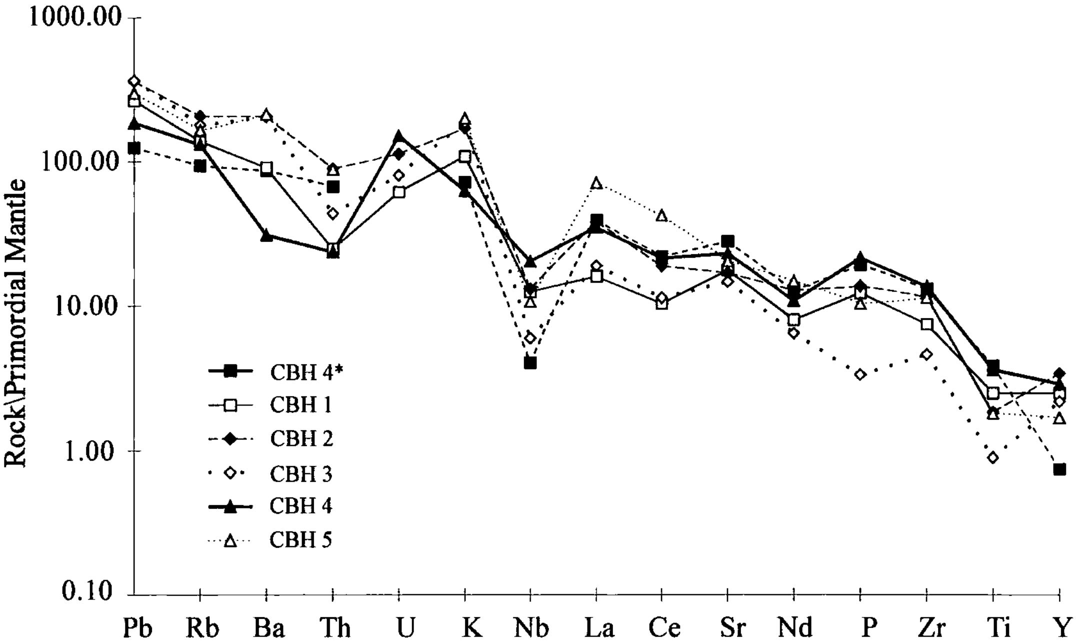
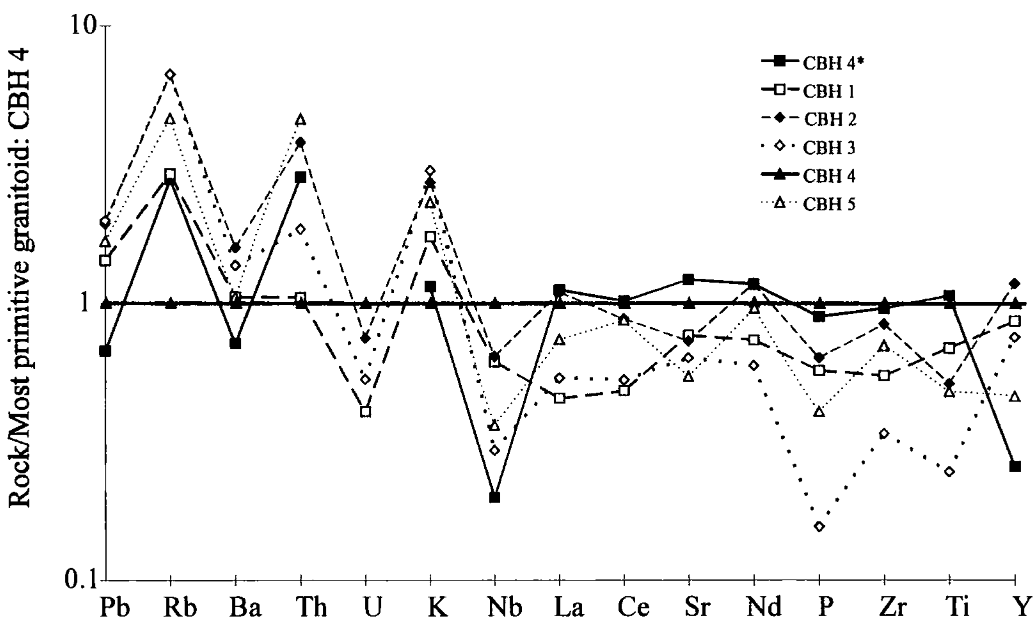


Figure 7:15:- Harker plots for the granitoids of Crobane Hill (wt. %)
(Horizontal axes (x) is SiO_2)



(A) Trace element variation within the "CBH" granitoids normalised to primordial mantle (after McDonough *et al.* (1992). P^{5+} from Sun (1980)).



(B) Trace elements variations within the "CBH" granitoids normalised to most primitive granitoid (CBH 4).

Figure 7:16:- Spider diagram plots showing trace element variation within the Crobane Hill granitoids.

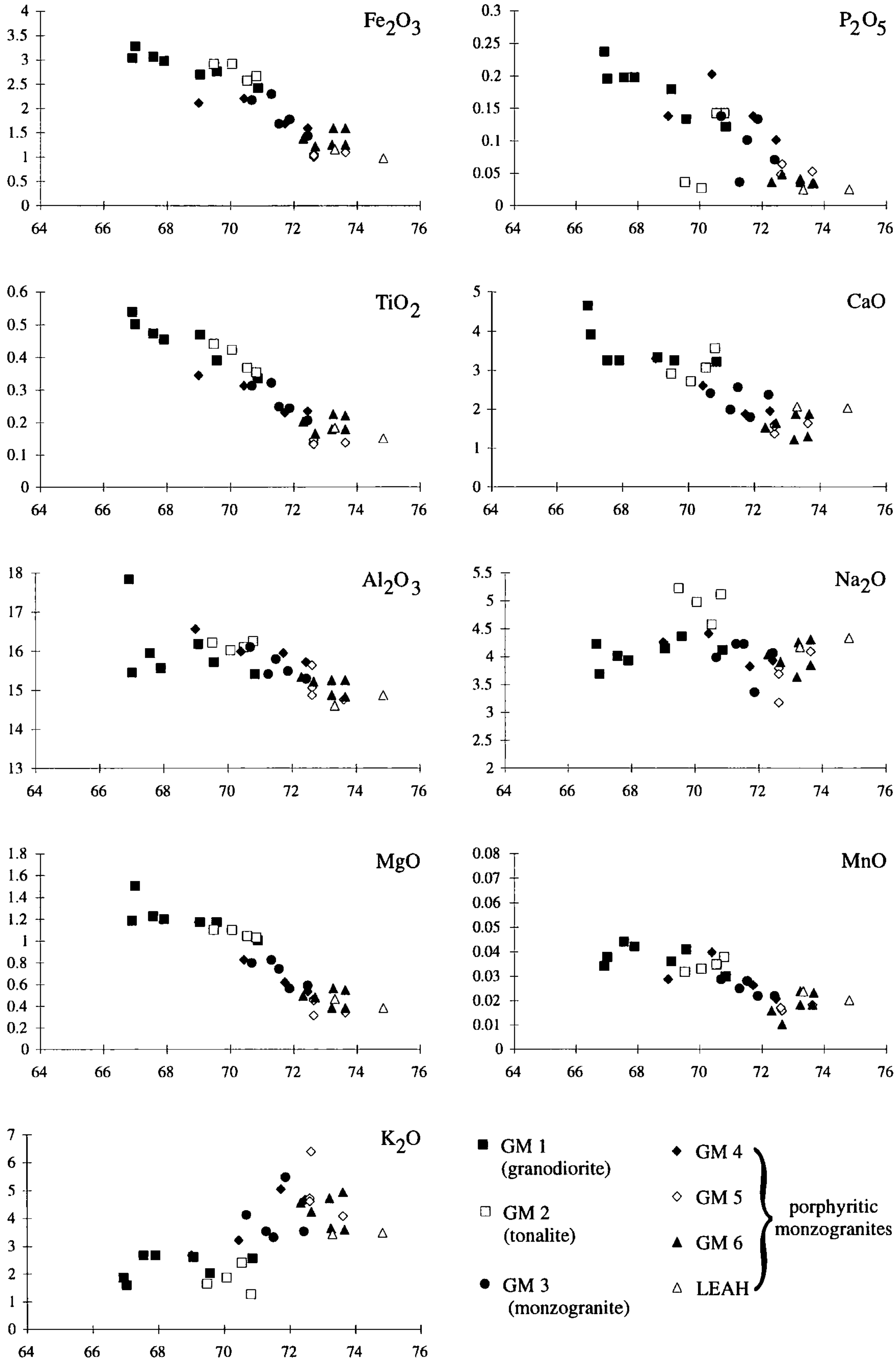
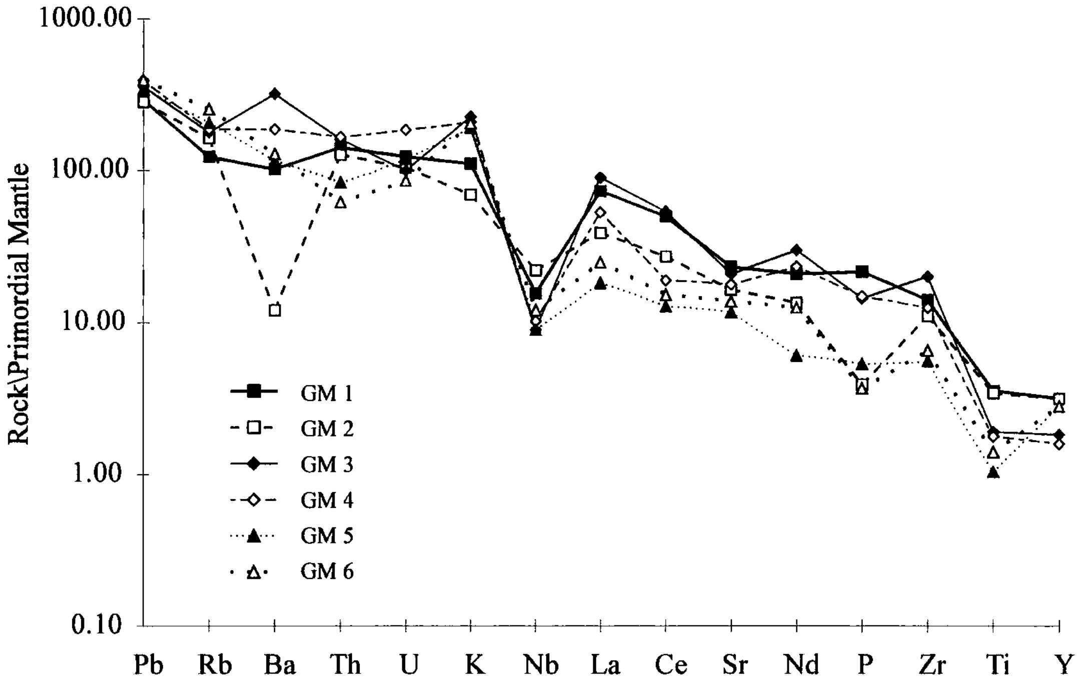
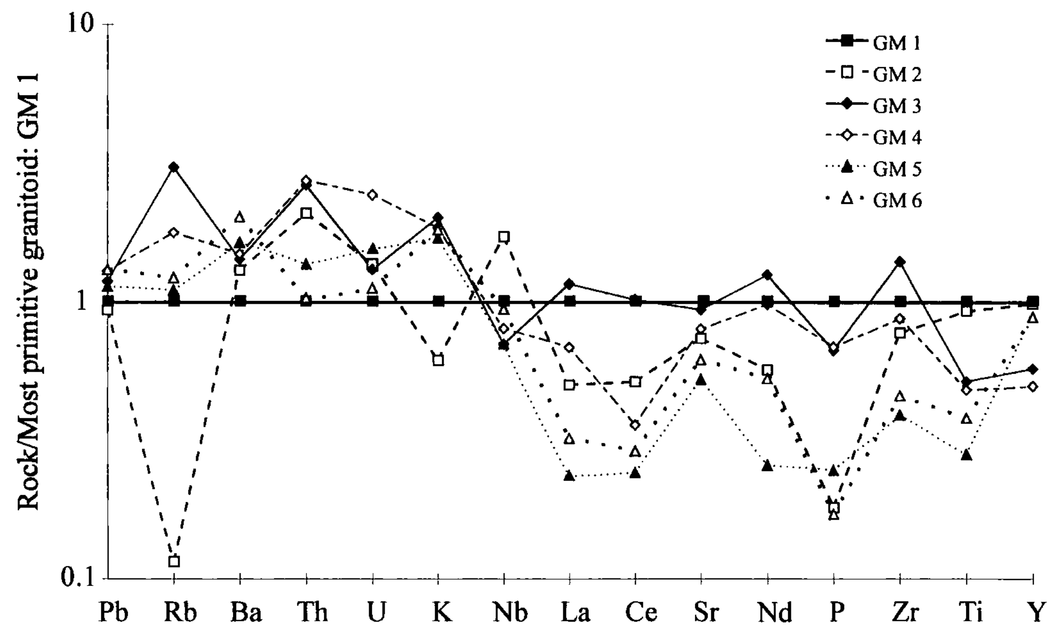


Figure 7:17:- Harker plots for the granitoids of the Glendowan Mountians (wt. %) (Horizontal axes (x) is SiO_2)



(A) Trace element variations within the "GM" granitoids normalised to primordial mantle (after McDonough *et al.* (1992). P^{5+} from Sun (1980)).



(B) Trace element variations within the "GM" granitoids normalised to the most primitive granitoid (GM 1).

Figure 7:18:- Spider diagram plots showing trace element variation within the granitoids of the Glendowan Mountains.

as this coarse, pink porphyritic monzogranite bears strong resemblance, in hand specimen, to the GM 5 monzogranite of the Glendowan Mountains area. As in the other two areas the granites of Glendowan Mountains have been plotted on Harker plots (figure 7:17) to show overall major element trends and to also whether or not these granitoids show any similarities to the other two areas. The overall geochemical trends show that the oldest granitoids are again the most primitive, with younger monzogranites becoming progressively more evolved. Again only K_2O shows a increase with increasing SiO_2. The only plot which shows any real deviation from a linear trend is Na_2O v SiO_2 where the GM 2 tonalite lies above the other granitoids. This feature is consistent with K-feldspar not being a fractionating phase during the formation of this granitoid.

The primordial mantle normalised plot (figure 7:18a) for the granitoids of this area is very similar to that seen in the Sruhanavarnis Valley. The GM 2 tonalite again shows a marked negative Ba anomaly, similar to the SRU 2 tonalites. Geochemical and petrographic evidence suggest that these tonalites are essentially the same. The GM 3 aphyric monzogranite which, from field relationships, was shown to be earlier than the porphyritic monzogranites shows quite distinct trace element distribution, tending to be relatively enriched in the HFSE. The abundances of these elements lie close to the GM 1 granodiorite trend (figure 7:18b). The significance of this granitoid is not certain, as it has not been seen in any of the other mapping areas within the Main Donegal Pluton.

7:4:2:4 Summary of the heterogeneous zones

The granitoids from the three mapping areas within the central regions of the pluton all show a similar pattern, where the oldest granitoids are the most primitive. The majority of this geochemical variation can be observed at the outcrop scale (i.e. grain size, petrography and texture). The only exception to this is the tonalites of the Sruhanavarnis Valley where the variation is more cryptic. In this area there are more silica-rich tonalites which contain more titanium-rich biotite, in contrast with more intermediate tonalites and their more iron-rich biotites. The tonalites and granodiorites from the three mapping areas in chapter 4 were plotted on Harker-type plots and also trace element variation diagrams (see figure 7:19 and 7:20), to see if there any similarities between the various phases. The tonalites of Galwollie Hill were also plotted on these diagrams. The following conclusions can be made from these diagrams:-

1) The most primitive of all the earlier granitoids is the SRU 1 granodiorite, as is revealed from several of the Harker plots, most notably Fe_2O_3, TiO_2 and P_2O_5

(figure 7:19). This feature is in very strong accordance with the field evidence, as it was believed that some of the oldest granitic material in the pluton would be found to

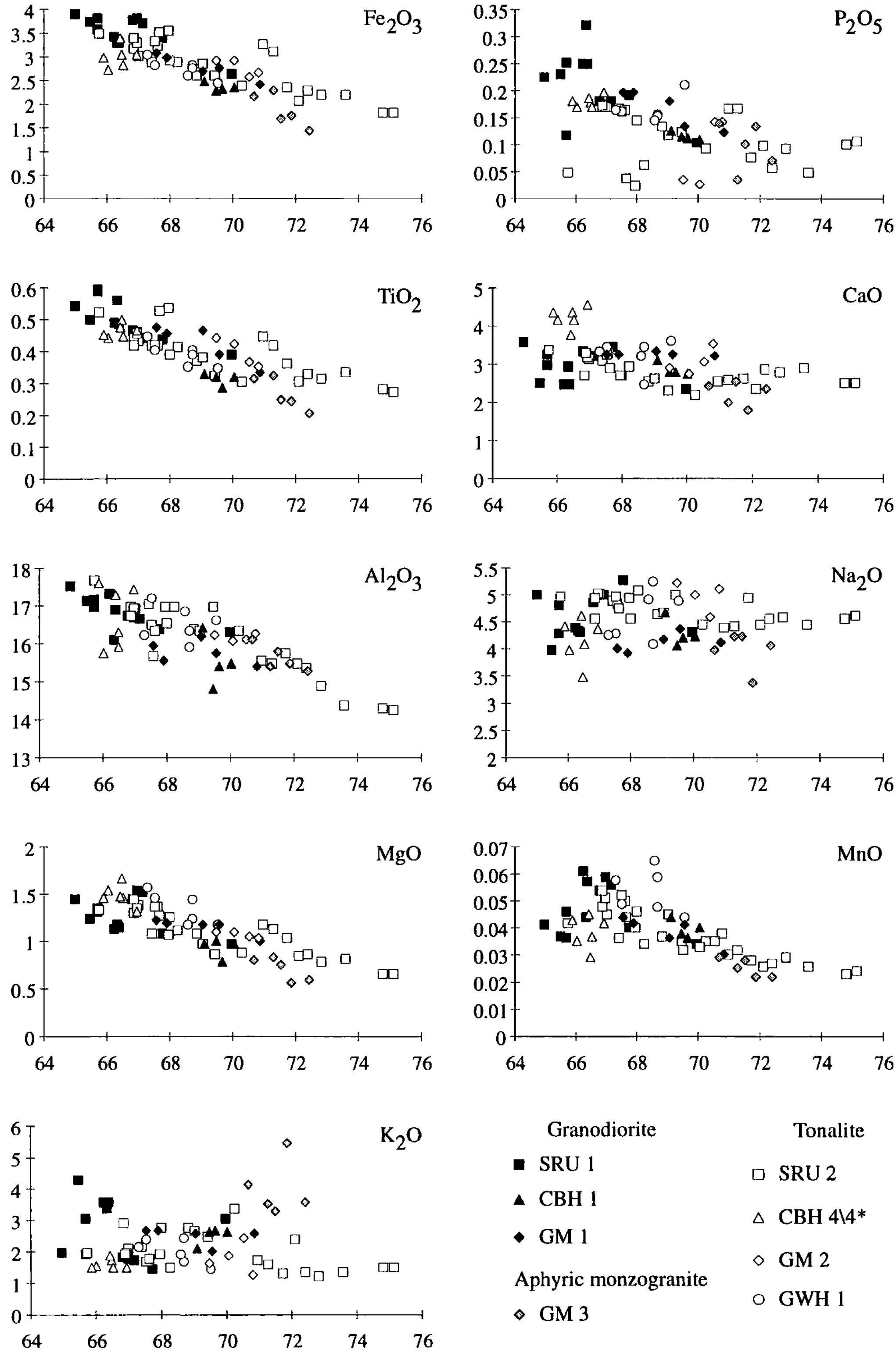


Figure 7:19:- Harker plots for the early granitoid phases seen within the heterogeneous granitoid zones of the Main Donegal Pluton (wt. %)
(Horizontal axes (x) is SiO_2)

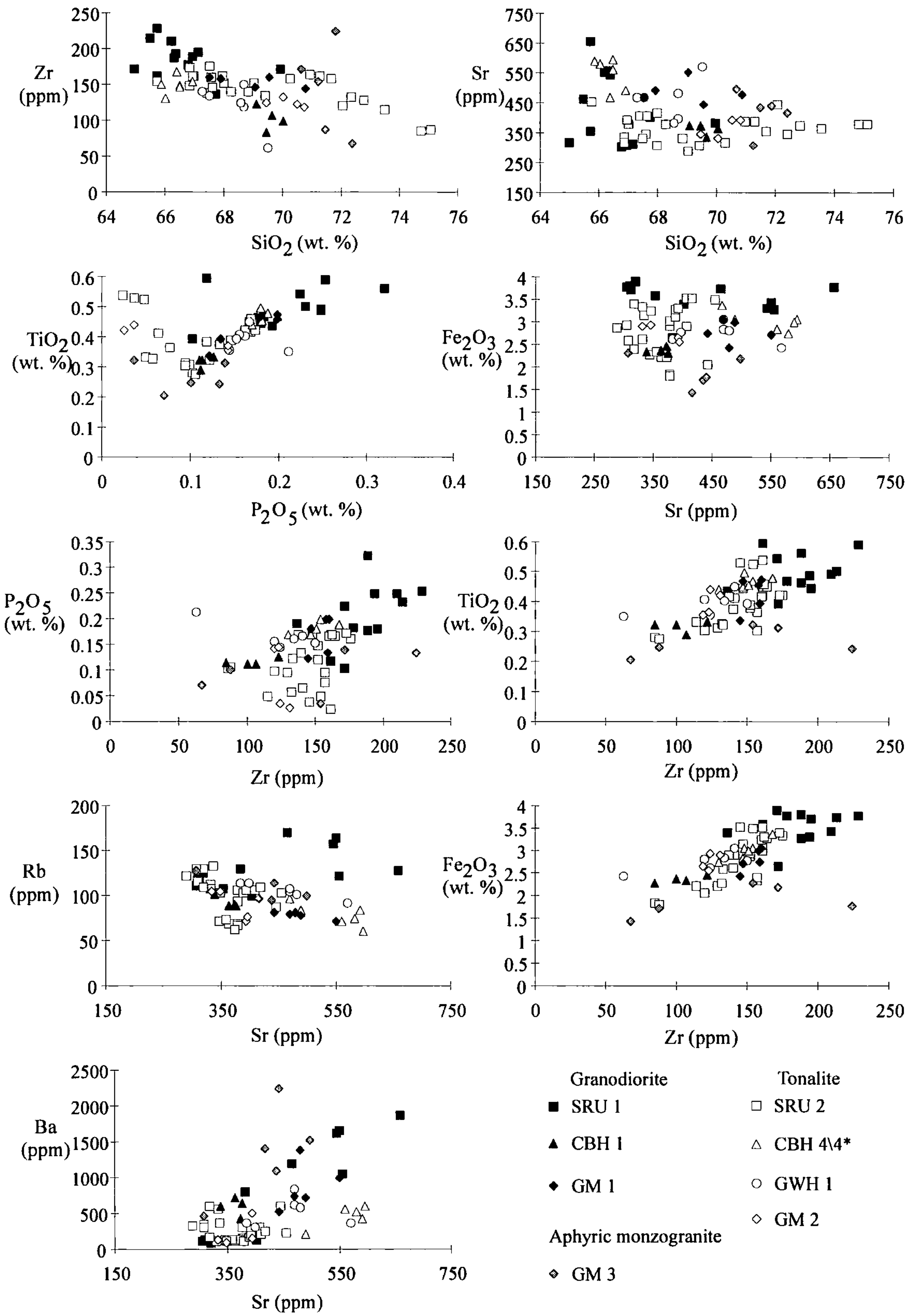


Figure 7:20:- Trace element variation diagrams for the early granitoid phases seen within the heterogeneous granitoid zones of the Main Donegal Pluton.

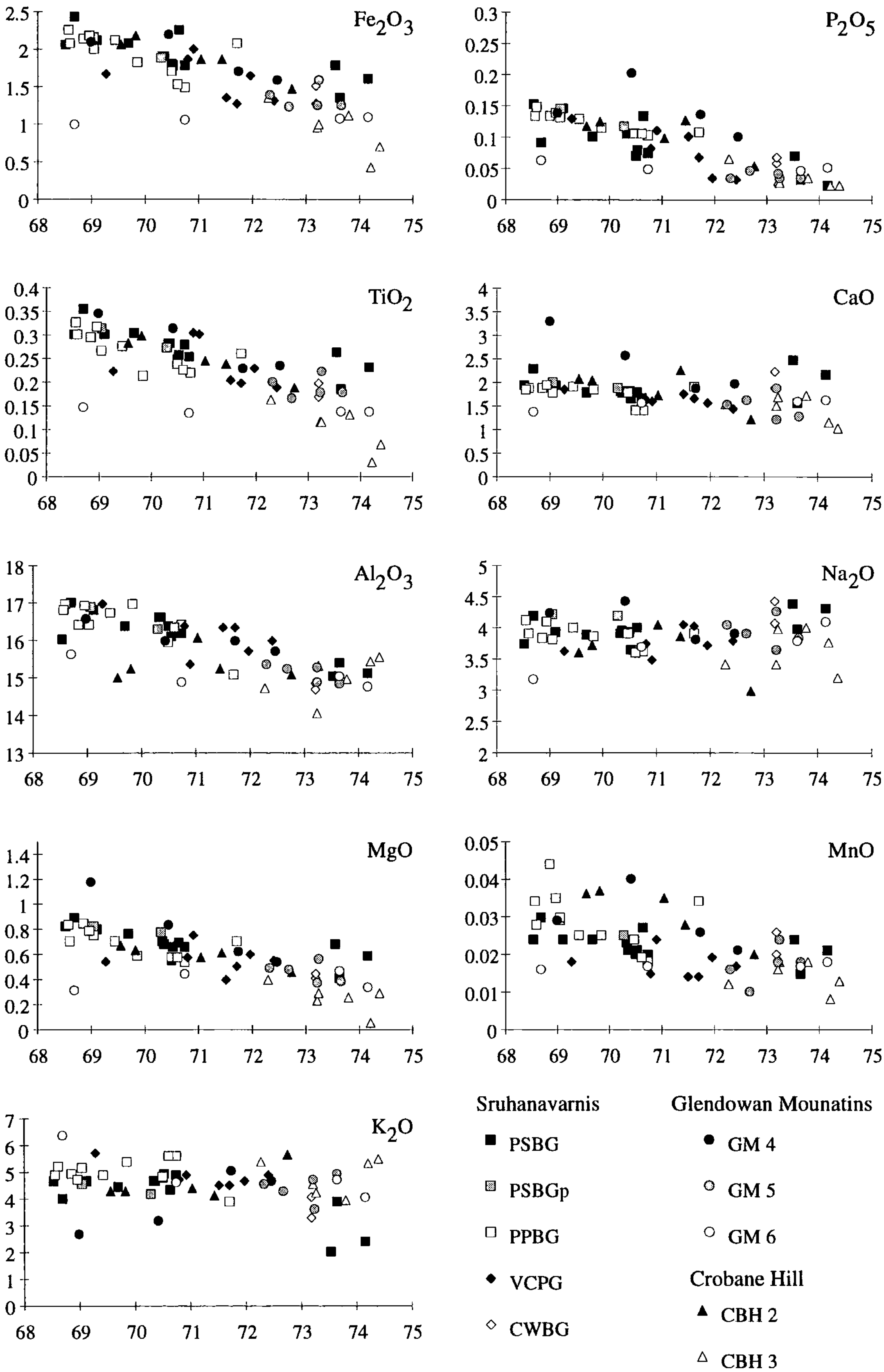


Figure 7:21:- Harker plots for the porphyritic monzogranites within the heterogeneous zones of the Main Donegal Pluton (wt. %)
(Horizontal axes (x) is SiO_2)

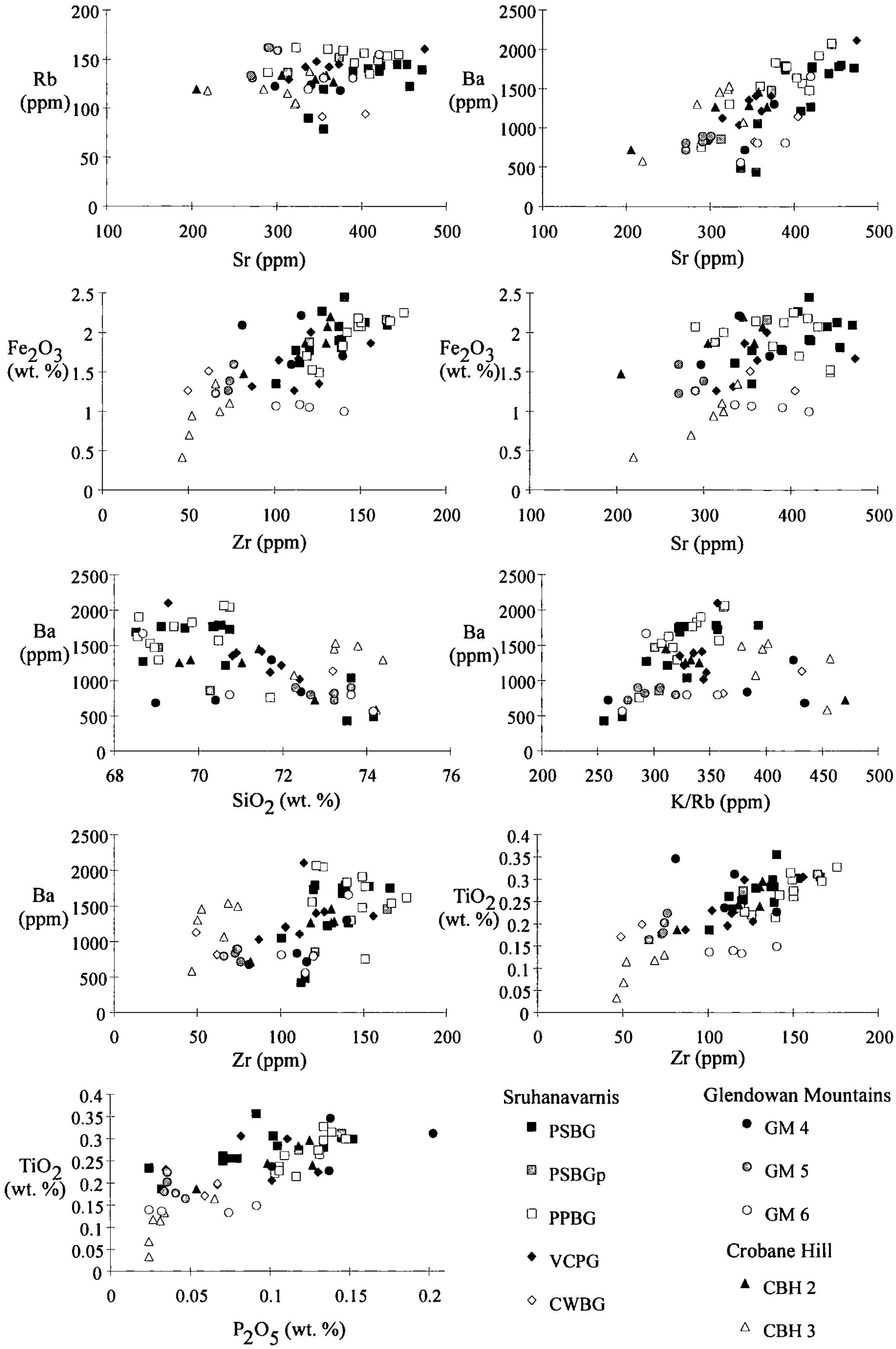


Figure 7:22:- Trace element variation diagrams for the porphyritic monzogranites within the heterogeneous zones of the Main Donegal Pluton.

either side, and to a lesser extent within, of countryrock material, i.e. raft-zones. Assuming that the most primitive granodiorites are the oldest (as feature seen in four detailed mapping areas) then the granodiorites of the Sruhanavarnis Valley may be some of the oldest granitoid material within the Main Donegal Granite.

2) The granodiorites from the three areas (i.e. SRU 1, GM 1 and CBH 2) are all quite distinct and show no strong correlation with one another, implying that this earlier highly disrupted early pulse was quite heterogeneous in character.

3) The tonalites show stronger similarities in both appearance and in geochemical composition suggesting overall greater homogeneity within this pulse.

The Harker and trace element plots (figure 7:21 and 7:22) of the monzogranites within the central areas of the pluton do not display any clear "clumping" which correlates with textural differences observed in the field (apart from the Rb-Sr plot) It was noted that in chapter 4 that the subtle differences within the porphyritic monzogranites may correspond to separate sheet-like pulses. Overall the medium-grained SRU 3 variants (PSBG, PSBGp and PPBG) and CBH 2 monzogranite tend to be the most primitive of the porphyritic monzogranites seen in the central regions of the pluton. The coarser porphyritic monzogranites of the Sruhanavarnis (VCPG and CWBG) and the Glendowan Mountains (GM 5 and GM 6) are more evolved. The CBH 3 aphyric monzogranite of Crobane Hill is the most evolved major granitoid phase seen within the central heterogeneous zones.

7:4:3 The geochemistry of Galwollie Hill "transitional boundary"

In the Galwollie Hill area the majority of the samples were collected from within the Map C area. Further samples of the Trawenagh Bay Granite have been obtained from the Ballynacarrick and Trusklieve Hill areas. The samples from this latter pluton are dominantly monzogranitic in composition.

The granitoids of this area were plotted on Harker diagrams (7:23) to see whether the general major element behaviour of these granitoids is related to the overall mineralogy. As seen in other areas all the oxides, apart from K_2O, decrease as the amount of SiO_2 increases. The following conclusions can be made:-

1) The GWH 4 monzogranite, which in Chapter 6 was believed to be more weakly deformed GWH 5, does appear to be the same granite because in all of the Harker plots these two monzogranites plot within similar areas. In figure 7:24a the primordial mantle normalised trace element spider diagram patterns for the GWH 4 and GWH 5 monzogranites are also identical and hence reinforce the major element geochemical data. It was this granitic phase which displayed peraluminous affinities when cationic proportions were calculated (i.e. $(K+Na+2Ca)/Al<1$).

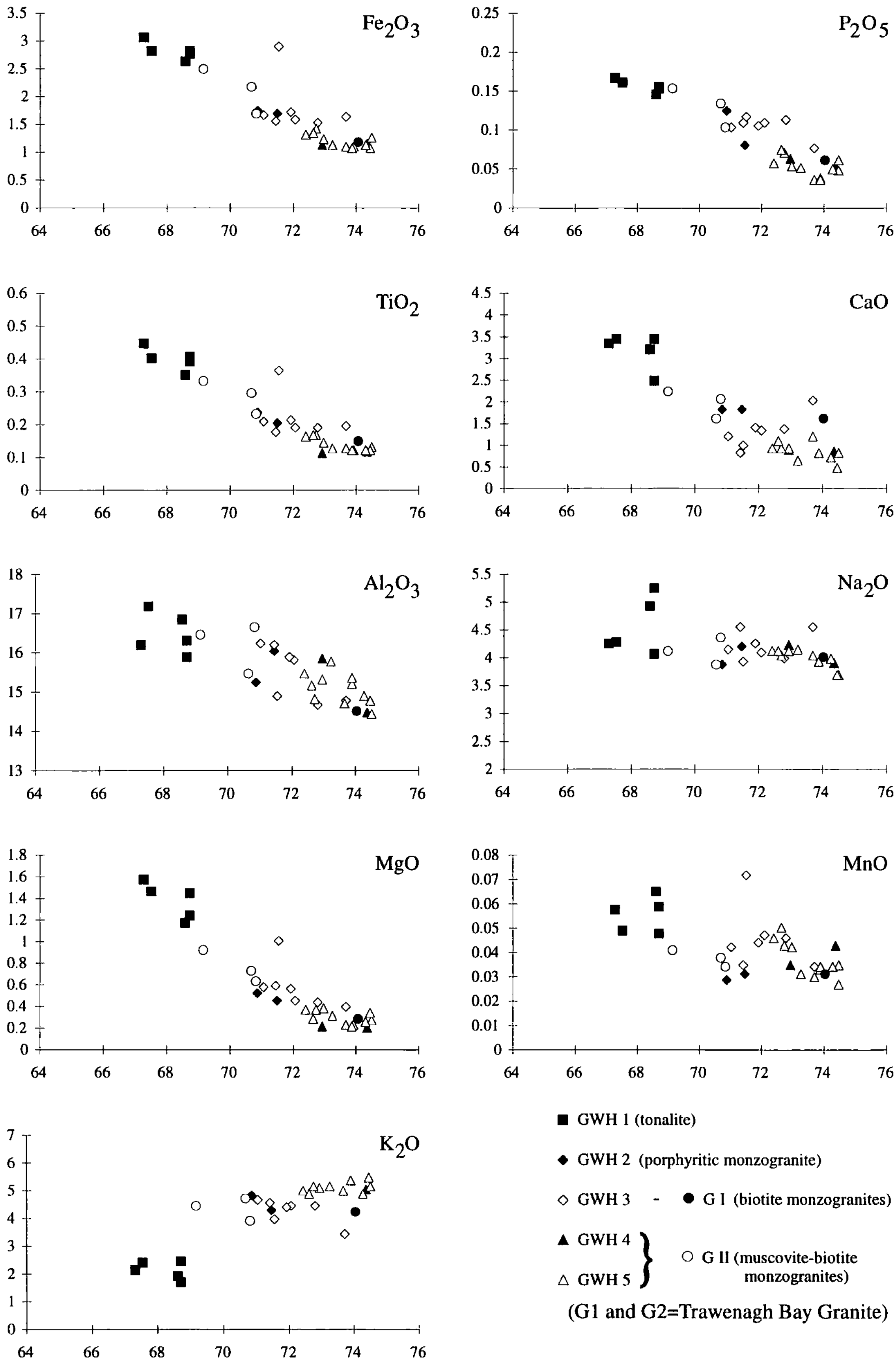
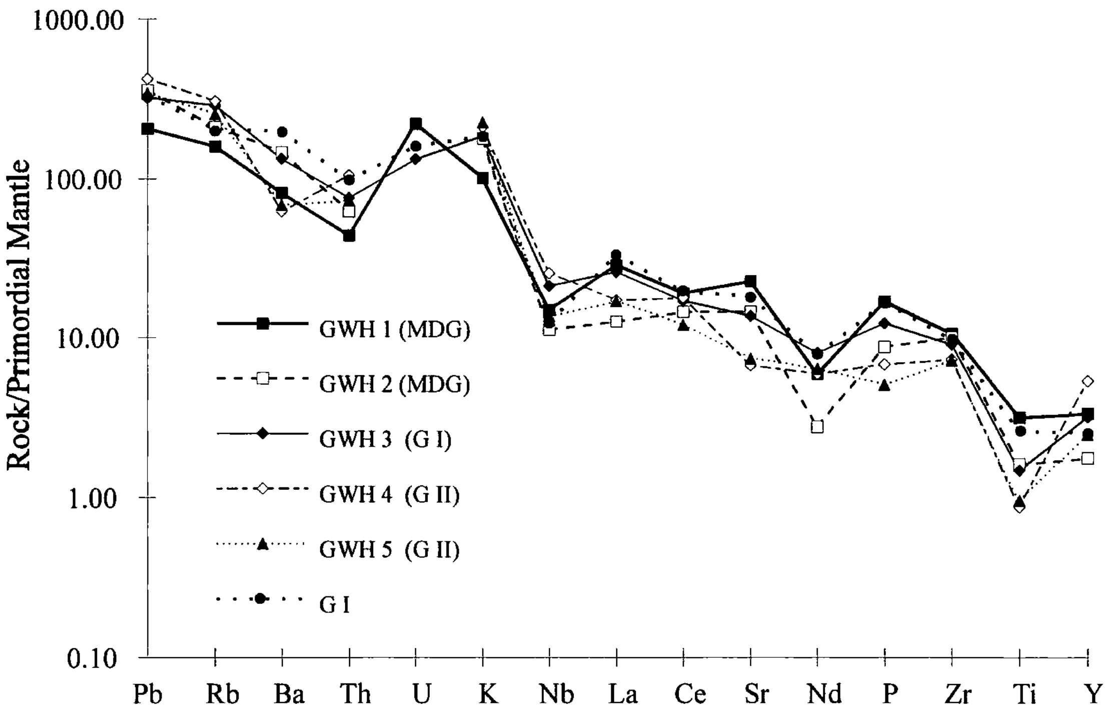
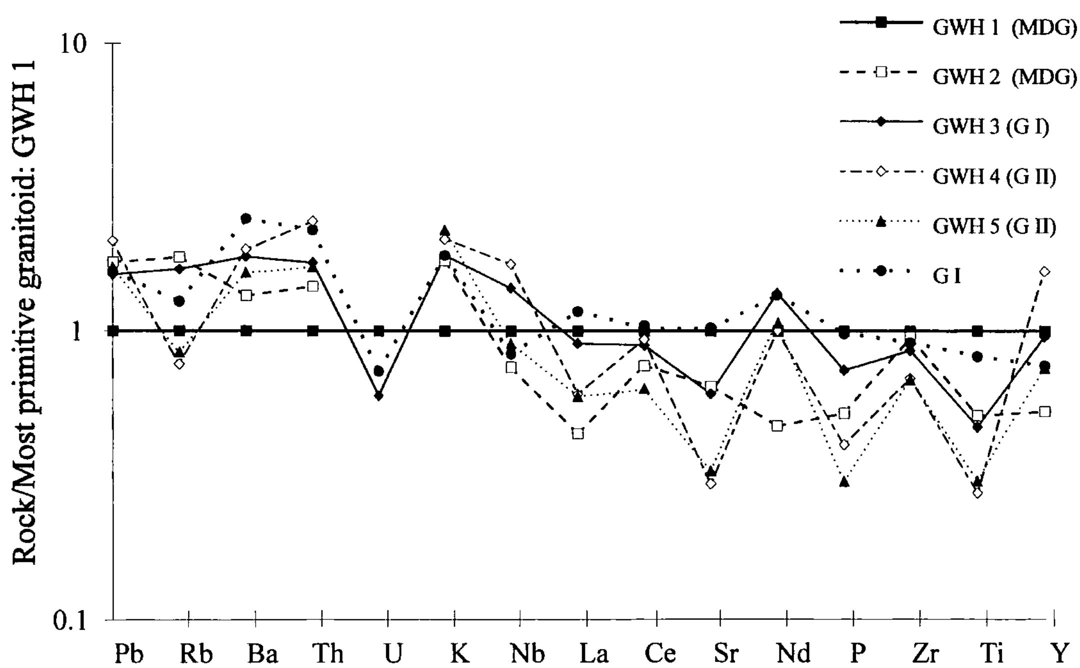


Figure 7:23:- Harker plots for the granitoids of Galwollie Hill (wt. %)
(Horizontal axes (x) is SiO_2)



(A) Trace element varaition within the "GWH" granitoids (and G I) normalised to primordial mantle (after McDonough *et al.* (1992). P^{5+} from Sun (1980)).



(B) Trace element variation within the "GWH" (and G I) granitoids mormalised to most primitive granitoid (GWH 1).

Figure 7:24:- Spider diagram plots showing trace element variation within the Galwollie Hill granitoids (MDG and TBG pluton members present).

No other major granitoid phase within the Main Donegal Granite displays such an affinity. The GWH 4 & 5 monzogranites belong to the G II pulse of the Trawenagh Bay Granite.

2) Apart from the samples obtained from the Trawenagh Bay Granite, the granitoids of Galwollie Hill again show a trend where the oldest tonalites are the most primitive with younger monzogranites becoming progressively more evolved.

In figure 7:24a the trace element data normalised to primordial mantle are very similar to the plots from the other areas, although subtle differences exist. The GWH 1 tonalite does not show the pronounced Ba negative anomaly that is also seen within the GM 2 and SRU 2 tonalites. The G I sample of Trawenagh Bay (GR 789028) shows similarities to the more deformed and reddened sample of GWH 3 obtained from the Map C on Galwollie Hill. In the spider diagram normalised to the most primitive granitoid of this area (GWH 1 tonalite) (figure 7:24b) the G I monzogranite is quite enriched in the HFSE. As was discussed in the petrographic section this monzogranite is quite rich in epidote with these crystals commonly having allanitic cores. Allanite is a mineral which commonly accepts trace and rare-earth elements into its lattice and hence probably accounts for the high levels of the HFSE within this granitis facies. In comparison with the other granitoids in this plot, the GWH 3 monzogranite also has relatively high levels of these elements but generally not as high as the G I sample.

7:4:4 The relationship of the central zones to the homogeneous zones

This section will address the variation seen in the Crockmore and the Binaniller apophyses and also the homogeneous porphyritic monzogranites in the north-western part of the pluton. These granitoids will be compared and contrasted with the petrographically similar monzogranites of the central zones. From field evidence it is believed that the homogeneous pink porphyritic monzogranites in the NW part of the pluton may be related to the abundant more variable porphyritic monzogranites that are present within the more central parts of the pluton. The relationship of the Binaniller apophysis may also be similar, although the exact relationship of the Crockmore apophysis to the porphyritic monzogranites is uncertain because this granioid is non-porphyritic. As was discussed in Chapter 5, the author views the Crockmore apophysis as one of the younger units within the Main Donegal pluton.

The monzogranites of the homogeneous zones were plotted against the porphyritic monzogranites seen within the central zones, to see if they show any similarity. Furthermore the younger aphyric CBH 3 monzogranite and the G I (GWH 3) and G II monzogranites (GWH 4& 5) of Galwollie Hill were also plotted.

In figures 7:25 and 7:26 various Harker plots are shown, as are certain trace element plots. The following conclusions are based on these plots are:-

i) The homogeneous pink porphyritic monzogranites are geochemically very similar to the monzogranites of the central areas which might imply they were penecontemperaneous during emplacement.

ii) There is no strong evidence to suggest that the Binaniller apophysis is geochemically distinct from the porphyritic monzogranites. In all plots the Binaniller monzogranites lie within the field of the other porphyritic monzogranites.

iv) The granites of the Crockmore apophysis are generally very homogeneous, with all the samples plotting within a well developed cluster. These granites can be differentiated from the other monzogranites by the relatively high values of strontium (see Rb-Sr and Sr-Ba plots).

v) The G II monzogranite of the Galwollie Hill area is generally quite evolved, as indicated by the Harker plots, tending to be rich in silica and depleted in most major oxides. The proportion of the K_2O and MnO (MnO levels are governed by the isomorphous replacement of Al^{3+} in the muscovite lattice) is slightly higher than the values seen within the Main Donegal monzogranites, although CaO and to a lesser extent MgO levels tend to be slightly lower. The lower values of CaO are also seen within some samples of the G I monzogranite. Despite being rich in K_2O, the G II monzogranite has quite low Ba levels in comparison with MDG monzogranites, although the Rb levels are considerably higher (see figure 7:26). This might suggest the source to the G II granite was more enriched in Rb than the source to the MDG monzogranites, with this element preferentially entering biotite, muscovite or K-feldspar rather than Ba.

vi) The CBH 3 monzogranite is one of the most evolved monzogranites within the Main Donegal pluton, having a high silica content together with more depletion in major oxides and some trace elements than the porphyritic monzogranites.

7:5 Summary

The detailed mapping and geochemical sampling of the Main Donegal pluton has shown that the oldest granitoids within the pluton are generally the most primitive, with the subsequent granitoid phases becoming more evolved. These relationships suggest that some form of fractional crystallisation may have been occurring within the source region to the pluton, either within the zone where they originated or between there and the level of their final emplacement. Large-scale partial melting, without modification of the liquids by subsequent processes can be ruled out because, if this process was occurring one would expect the oldest granitoids to be the most evolved. This is because the minerals with lower melting temperatures

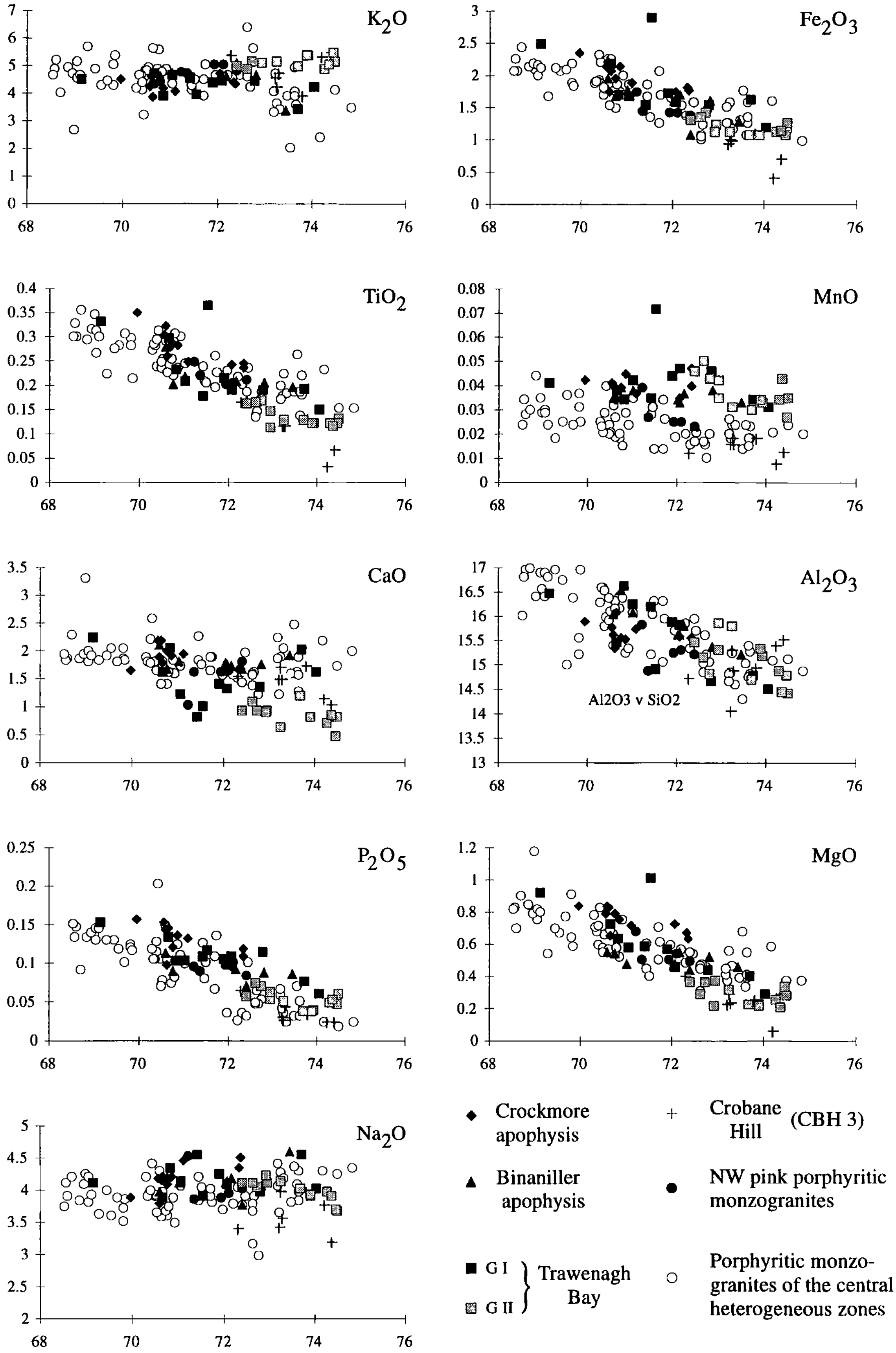


Figure 7:25:- Harker plots for the porphyritic monzogranites and younger aphyric monzogranites within the Main Donegal pluton (wt. %)
(Horizontal axes (x) is SiO_2)

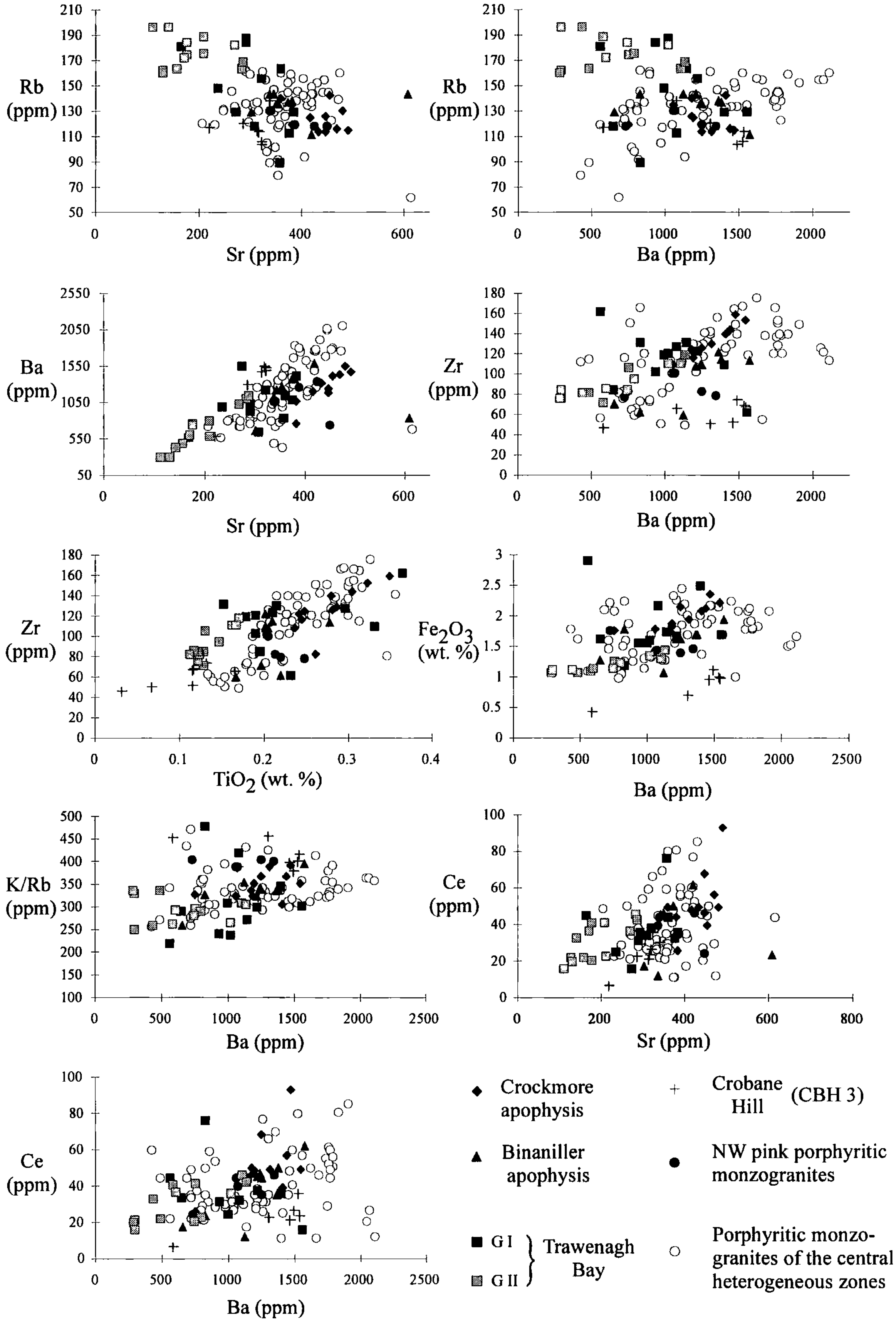


Figure 7:26:- Trace element variation within the porphyritic monzogranites and younger aphyric monzogranites within the Main Donegal pluton.

(i.e. quartz and alkali feldspar) will start to fuse first and hence younger granitoids become more primitive as either temperatures rise (allowing higher temperature mineral phases to melt) or general depletion in the source area of low-temperature melts. The field relationships imply that the pluton was constructed episodically; initially of biotite-rich granodiorite and then biotite tonalite and finally the voluminous porphyritic monzogranites, which generally show very subtle petrographic, textural and geochemical variations. These three major granitoid phases themselves show considerable internal variation, which might suggest that fractionation processes may have been occurring *within* these three groups. Although there is a tendency for the major granitoid phases to become more evolved, there are examples of minor intrusives which cross-cut all the main granitoids and are sometimes more primitive than their hosts (e.g. the CBH 4 tonalite dyke and similar cross-cutting tonalite dykes seen in the Moylenanav area). This implies that more primitive melts were still present, even at a late stage in the construction of the Main Donegal pluton. Therefore there are two options which might account for the origin of the different granitoids within the Main Donegal pluton:-

- The various granitoid components derived from several different source types during its evolution. The angular cross-cutting relationships of earlier phases relative to later ones implies the pluton was not constructed during one main event.
- Alternatively the three major granitoid phases may all be related to the same source but changing physical features in the source (i.e. increasing temperatures due to crustal thickening or mafic underplating of the lower crust and also the influence of major tectonic structures) over long periods of time may lead to a change in character of the granitoids being produced.

The absence of REE and isotope data limits the extent to which such problems can be investigated.

The G I pulse of the Trawenagh Bay Granite is compositionally very similar to the MDG porphyritic monzogranites. The G II pulse appears to be considerably more evolved than any other major phase of the Main Donegal pluton. The field relationship of the G II monzogranite on Galwollie Hill clearly implies it is younger than the Main Donegal Granite, its composition is also quite distinct. The chemistry of the smaller G III pulse is uncertain because this author has no samples.

Within the Donegal Batholith itself there is an overall trend for the earlier plutons to be more intermediate in composition (i.e. Thorr, Fanad, Ardara and Toories), whilst the later plutons become more acidic (i.e. Rosses, Main Donegal, Trawenagh Bay and Barnesmore). This relationship implies that the whole source region to the batholith became more evolved with time. Internally all of these plutons show normal zoning, where the earliest phases are compositionally the most primitive and subsequent phases become progressively more evolved. This invites the question of where in the crust this diversification of granitoid magmas was occurring. Was it in the source region or at storage areas (i.e. tectonic low-pressure zones) in the intervening crust, between the source and final level of emplacement? Such questions are beyond the scope of this thesis but imply that the source regions to granites may be very complex.

The final chapter will discuss the internal features of the Main Donegal Granite, in regards to its overall construction, and how it fits into the overall tectonic framework of NW Donegal.

Chapter 8

The emplacement and deformation of the Main Donegal Granite

8:1 Introduction

The aim of this chapter is to develop a model for the emplacement of the Main Donegal Granite (MDG). First, the influence of the original configuration of the Dalradian country rocks and older plutons of the Donegal Batholith before the emplacement of the MDG shall be discussed. This will be followed by a model proposing the emplacement mechanism and subsequent deformation of this pluton determined from the detailed field examination of this intrusion. Detailed mapping has shown that the appearance of sheeting within some areas of the pluton is not that obvious due to later sheets breaking up earlier formed granitoids which are now preserved as extensive zones of rafts within younger monzogranites which creates difficulty in identifying the boundaries of these later sheets. In other areas the sheeted nature is relatively obvious due to the presence of countryrock metasediments to either side of relatively homogeneous monzogranite. The factors which control the morphology and appearance of granite sheets will also be discussed in this chapter. The fundamental features of this emplacement model will then be compared and contrasted with other granitic plutons that have been emplaced within similar transcurrent shear zone regimes. The emplacement of granite plutons within shear zones has lead to the speculation as to the rate that granitic magmas are emplaced, and to what extent tectonic influences/forces control the ascent. Such issues will also be addressed in this chapter.

8:2 The emplacement model for the Main Donegal Granite

The current model for the emplacement of the Main Donegal Granite was put forward by Hutton (1982), who proposed the pluton was emplaced within a sinistral shear zone. Hutton (1982) stated that the pluton was essentially accommodated within the shear zone due to the development of a displacement gradient along the length of this structure which caused it to split axially, allowing granitic magma to be drawn into the shear zone. The present results of fieldwork in the present study has shown that the Main Donegal Granite has been subjected to syn-plutonic deformation related to a sinistral shear zone although the present work has allowed this model to be further developed and refined.

The following sections will develop a model for the incremental emplacement of the Main Donegal Granite based mainly from observations within the pluton.

8:2:1 The relationship of older plutons to the Main Donegal Granite Shear Zone (MDGSZ)

A study of the older plutons (Thorr and Ardara), which are now preserved as fragments within the present outcrop of the Main Donegal Granite, has provided evidence to show that the Main Donegal Granite Shear Zone (MDGSZ) was active prior to the emplacement of the Main Granite itself. At Carbat Gap xenoliths of the G I pulse of the Ardara pluton are preserved. It was noted that these xenoliths were more highly deformed than the host MDG facies which suggests they have been entrained within the shear zone for a longer duration. The geometry and overall appearance of the deformation fabric within these xenoliths is similar to that seen in the adjacent MDG facies, albeit weaker in the latter. Furthermore, the preservation of country-rock contacts with these xenoliths implies that this now xenolithic mass was dominantly autochthonous, and hence the shear zone was active in this area prior to the emplacement of the Main Donegal Granite. The "stalk" of the Ardara pluton may be a remnant of this G I prolongation of the Ardara pluton which intruded into the MDGSZ.

In the case of the Thorr pluton the evidence from xenolith distribution in the Main Donegal Granite suggests that a similar and somewhat bigger prolongation originally existed which also trended in a NE-SW direction. It is the present authors belief that this Thorr prolongation was also emplaced within the shear zone. In figure 8:1 the pre-Main Donegal Granite geology is shown illustrating the former extent of these plutons. The NNE trending concealed lineament of Hutton & Alsop (1996) is incorporated into this model, as these authors believed that the emplacement of the bulk of the presently exposed Thorr pluton was controlled by the lineament which they interpreted as a weak sinistral transtensional structure. Along the eastern contact of the Thorr pluton, Hutton & Alsop (1996) and McErlean (1993) observed evidence for sinistral syn-magmatic shearing and overprinting NNE trending S-C fabrics which Hutton & Alsop (1996) attributed to minor displacements along the lineament. The presence of these fabrics and the Thorr prolongation may imply that both the MDGSZ and the lineament may have been active contemporaneously during the emplacement

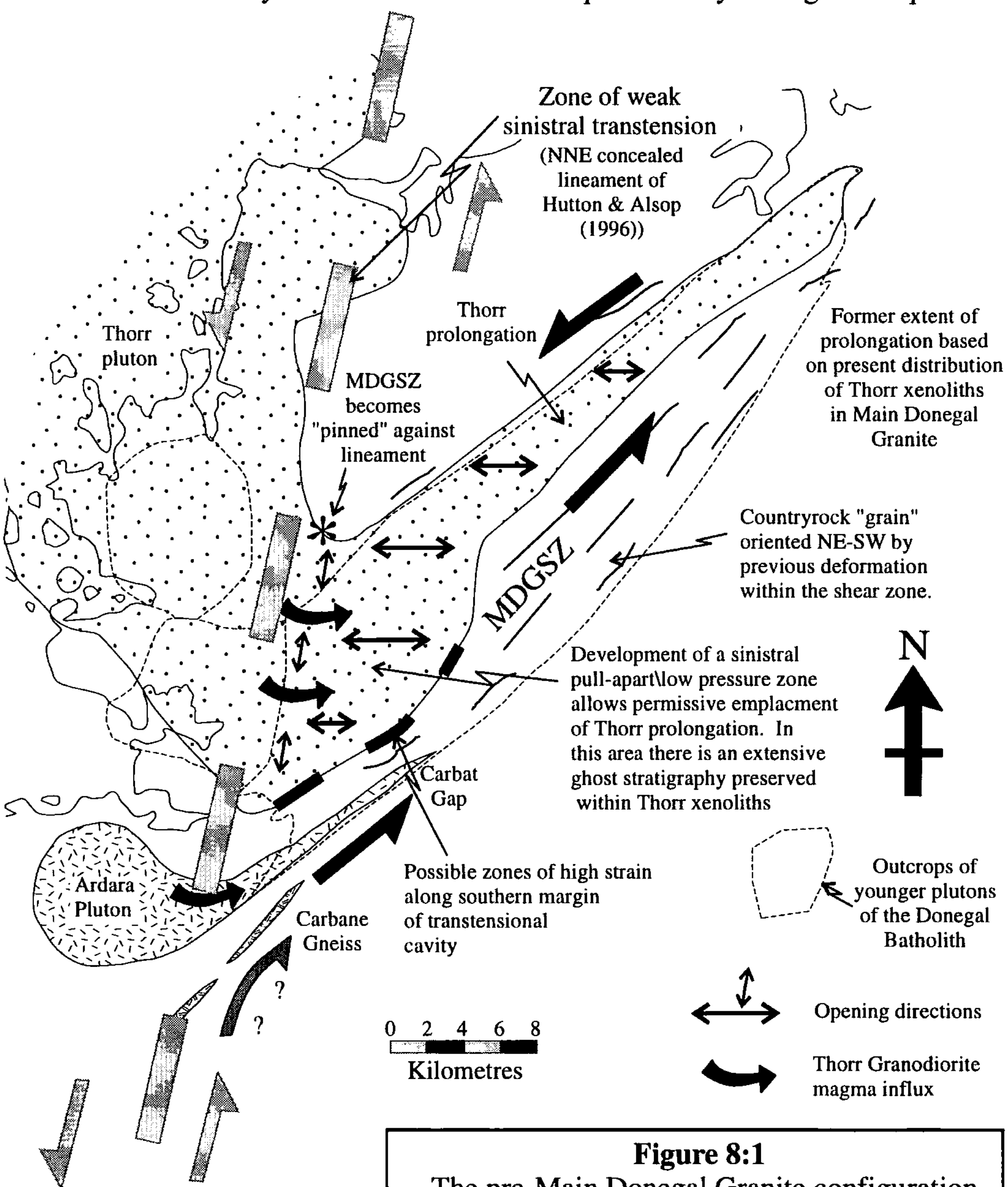


Figure 8:1
The pre-Main Donegal Granite configuration of the Thorr and Ardara plutons

of the Thorr pluton (see later on). Hutton & Alsop (1996) also stated that the Ardara pluton may have been sited at the intersection of this NNE lineament with the NE-SW trending MDGSZ. These authors believe that displacements along Donegal lineaments are very minor (<10 Km) as the Dalradian stratigraphy shows no major offsets to either side of this feature (see figure 8:1 for its believed occurrence). These authors and others (e.g. Jacques & Reavy (1994) claim that such lineaments may not accommodate much displacement, but more importantly they form anisotropies within the basement which may be exploited by igneous melts during subsequent magmatic events. Hutton (*pers comm.* 1997) has stated that smaller, parallel lineamental features may lie to either side of the main lineament and consequently may have exerted far more subtle controls on features seen at the present surface. The possible occurrence of these more subtle basement features will be considered in following sections.

The xenolithic remnants of the Thorr prolongation are relatively undeformed, a feature in strong contrast to the xenoliths of the Ardara pluton at Carbat Gap. For this reason it is believed that the Thorr prolongation was essentially accommodated by the development of a combined low pressure zone and transtensional cavity resulting from interference between the MDGSZ and the NNE lineament. This may have been due to the NW part of the MDGSZ becoming "pinned" (the present Crockator area) against the lineament which would create a localised transtensional cavity in this area (see figure 8:1). When studying the distribution of country rock xenoliths within the Thorr pluton the majority of them occur within the region to the south and south-east of Crockator which may confirm the existence of a transtensional low pressure zone. The presence of these xenoliths may relate to tectonic tensile fracturing resulting from dilation plus some hydraulic fracturing by the invading Thorr magma itself. The abundance of country rock xenoliths within this zone suggest that displacements within this transtensional cavity need not have been that great. Figure 8:1 shows how the orientation of arrows (the east-west arrows) is broadly consistent with the extension direction predicted within the transtensional lineament and also compatible with the extension direction of the MDGSZ (Sanderson and Marchini 1984). Although essentially passively emplaced, the lateral confining pressure of country rocks and magmatic pressure (buoyancy and "head") may have allowed the "wedging-apart" of the Dalradian stratigraphy in a NE direction. The extent to which the MDGSZ was active is uncertain, as the strains along the NW margin of the prolongation are very high, relating to later movements along the MDGSZ during the construction of the Main Donegal Granite. Along the southern margin of this prolongation, (the Derryveagh 2 raft-zone), which is located further away from marginal deformation associated with the emplacement of the Main Donegal Granite, one might expect some degree of preservation of earlier MDGSZ fabrics. In the Newbridge area of the Barnes Gap, along the Newbridge-Glen road, the deformation within the Derryveagh 2 raft-zone is intense and this may represent earlier deformation along the southern margin of the Thorr prolongation. It was noted that within the MDG facies, to the immediate SW, the strain becomes less intense. This, therefore, may imply that the southern margin to the Thorr prolongation was more highly deformed than the more medial regions of the prolongation. This is similar to the present marginal areas of the Main Donegal Granite which are more heavily deformed than the central areas i.e. strains may localise along a viscosity contrast e.g. granite-country rock contact. This author heeds caution, as in some parts of the Main Donegal Granite certain raft-zones have localised strain during the cooling of the pluton, due to the granite becoming rheologically stronger than the meta-sediments, with strain progressively localising into weaker pelitic material. It is uncertain whether or not the width of the prolongation illustrated in figure 8:1 is the true width, as it has now been intruded by the younger homogeneous MDG pink porphyritic monzogranites. The options are as follows:-

- If this was the true width of the prolongation, then the later porphyritic monzogranites would have had to have been completely passively emplaced, i.e. the volume of missing Thorr Granodiorite was replaced with a similar volume of monzogranites. This invokes mechanisms such as large-scale stoping or other permissive mechanisms. If this were the case, then the lack of evidence for the original Thorr material needs to be explained. There is the possibility that it has been eroded away although the very homogeneous nature of the pink porphyritic monzogranites makes this seem unlikely.
- The Derryveagh 2 raft-zone was originally in close proximity to the autochthonous strip of Thorr currently preserved along the NW margin of the Main Donegal pluton, i.e. the prolongation was considerably narrower. This option implies that the pink porphyritic monzogranites essentially "wedged" apart (or were accommodated by space-creating displacements along the MDGSZ) the Thorr prolongation.
- A possible combination of the above options i.e. the prolongation, for example, may have been half the width in figure 8:1 with the porphyritic monzogranites having been emplaced by a combination of wedging and volume for volume replacement mechanisms .

The author prefers the last option, although it is very difficult to prove which option is correct. In figure 8:1 the width of the prolongation is illustrated with reference to the distribution of xenoliths within the Main Donegal Granite and to the present outcrop margins of this latter pluton.

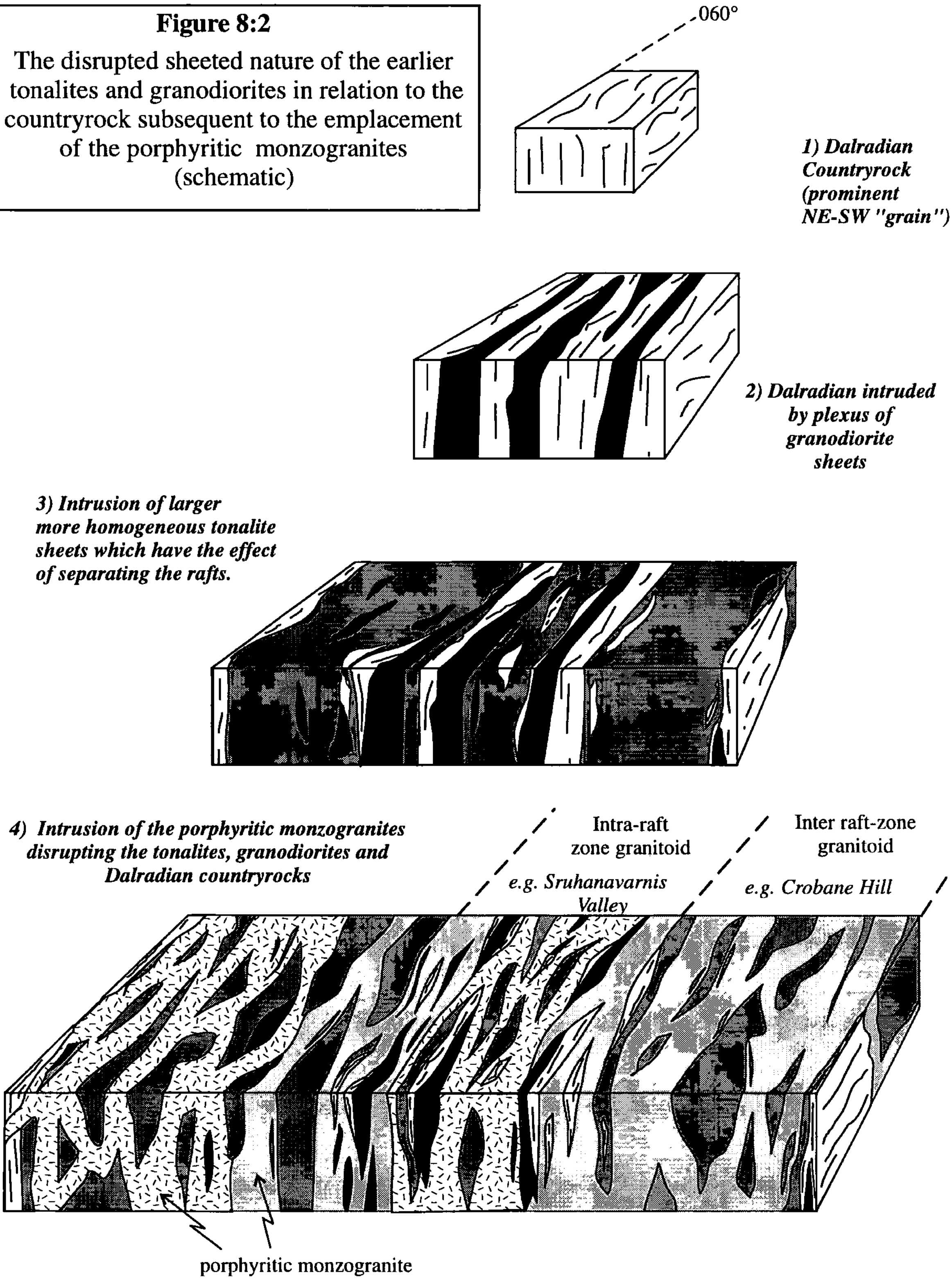
At the Carbat Gap, the Ardara xenoliths are more intensely deformed than the Thorr xenoliths. This suggests that in this area of the shear zone, strains were much higher. In figure 8:1 the displacement associated with the transtensional cavity opening for the Thorr prolongation would have to be accommodated along the southern margin of this cavity, i.e. the present Cleengort Hill-Croaghleheen area with this possibly accounting for the high strain recorded in the xenoliths (although this implies the Ardara pluton is older). Removal of the MDG facies may result in the Tonalite Xenolith Zone (TXZ) of Carbat Gap being closer to this margin and hence the high strains may have resulted from the accommodation of deformation associated with the strain related to the formation of the transtensional cavity discussed above. The Ardara prolongation may have been forcefully emplaced along the shear zone (i.e. not accommodated within a low pressure area of the shear zone) into the metasediments of this area and was being actively deformed during its emplacement. The divergence of the pelite rafts at the NE end of the xenolith zone attests to forceful emplacement by lateral wedging of the host country rocks. In the main Ardara pluton itself there is clear evidence from the granite and the country rock to suggest that forces associated with emplacement of the pluton were easily sufficient to push aside its country rocks. In chapter 6 it was noted that the Ardara xenoliths at Carbat Gap do show weak plagioclase PFC fabrics which are parallel to the foliation in the xenoliths, despite CPS strains being very high. This alignment of plagioclase, and the coplanar overprinting nature of the foliation is therefore, possible evidence for the syn-kinematic nature of the Ardara prolongation in regards to the MDGSZ.

In the Glenties area Hutton (1982) reported that the orientation of the MDGSZ rapidly changes to a more NNE-SSW orientation. This feature may be related to possible lineamental control with the strain being dissipated along it. The deformation of the Carbane Gneiss may be related to the MDGSZ changing orientation to a more NE -SW direction. In this area, some displacement along the lineament may have been accommodated along the MDGSZ (see figure 8:1). It was observed that the Carbane Gneiss has been subjected to high-temperature solid-state deformation which may relate to earlier movements along the shear zone before the emplacement of the Main Donegal Granite.

8:2:2 The initiation of Main Donegal Granite emplacement

The evidence from the previous section implies that the MDGSZ was in operation before the emplacement of the Main Donegal Granite itself. Therefore one might expect this structure to have had an influence during the emplacement of the Main Donegal Granite. The earliest phase of intrusion related to the MDG is that of the fine to medium-grained, biotite-rich granodiorites which are now currently preserved in the Sruhanavarnis Valley, Croaghacullin area of the Glendowan Mountains and the summit regions of Farscallop and Mount Kinnaveagh. The medium-grained granodiorites of Crobane Hill (CBH 1) may be slightly younger than the more finer-grained varieties as in appearance and from outcrop relationships they show greater similarity with the medium-grained tonalites. In chapter 7 geochemical data suggested that these granodiorites were compositionally quite variable although the most primitive granodiorite occurred within the Sruhanavarnis Valley. It is the authors belief that MDG construction initiated in the central regions of the pluton along the old southern margin of the Thorr pluton, as no heterogeneous granites are seen to the NW of the Derryveagh 2 raft-zone (for the possible relationship of the marginal facies granitoids to the central portion heterogeneous zones, see later on). It was this early granodiorite which intruded into the Dalradian metasediments. It was noted in chapter 4 that these granodiorites were in the near vicinity of the Glenveagh 3A raft-zone. When the younger tonalites and monzogranites are removed from the NW end of the Sruhanavarnis valley this raft zone and associated granodiorites may well have formed the immediate envelope to the SE of the Thorr prolongation (the Derryveagh 2 raft-zone), assuming the space for the granites has been created. The true extent of this earlier granodioritic pulse is uncertain due to it having been totally disrupted by the emplacement of subsequent tonalites and the more voluminous porphyritic monzogranite. Figure 8:2 is a schematic block diagram which helps to illustrate the highly disrupted nature of these early granodioritic pulses. It is believed that these granodiorites were dominantly intruded as a plexus of sheet-like masses (possibly between 10-30 metres wide) within the Dalradian metasediments. In none of the mapping areas are any of these early sheets still preserved in their entirety. The fine-grained nature of these granodiorites in the Glendowan Mountains implies that they may have been chilled quite rapidly after emplacement into the metasediments. The coarser grain size of later major phases (tonalites and monzogranites) may relate to rising ambient temperature of the envelope associated with the intrusion of earlier phases. In the central zones where the strain is relatively low, it appears that these early sheets were intruded parallel to the present foliation and have not been rotated by any larger degree, towards the shear plane. On Crobane Hill the CBH 1 granodiorite is preserved as lensoid rafts within the younger monzogranites and all are aligned parallel to the foliation. Again, the strain within this area is minimal and it appears that these earlier granodiorites were fractured in this orientation, suggesting some external tectonic control i.e. movements along the shear zone during the emplacement of later phases. The orientation of the sheets is not believed to be directly related to the MDGSZ as during simple shear one expects the sheets to initiate at 45° to the shear zone boundary. Subsequent deformation would cause these earlier sheets to be rotated to form a sigmoidal shape (Ramsay & Graham 1967). This sigmoidal sheet-like geometry is not seen within the Main Donegal pluton. In a transtensional situation the extension direction in relation to the finite strain ellipsoid is at an even higher angle to the shear zone boundary (Sanderson & Marchini 1984). In the case of the Main Donegal Granite, it is interpreted that the orientation of these earlier granodiorite dykes has been influenced by the pre-existing geometrical

Figure 8:2
The disrupted sheeted nature of the earlier tonalites and granodiorites in relation to the countryrock subsequent to the emplacement of the porphyritic monzogranites (schematic)



structure of the country rock. This pre-existing "grain" in the country rock may be related to earlier deformation along the shear zone before the emplacement of the Main Donegal Granite.

The medium-grained tonalites were the next phase of intrusion, being emplaced predominantly into the earlier granodiorites and country rock metasediments. The similarity of these tonalites, especially in the Sruhanavarnis Valley and the Glendowan Mountains implies they may have formed larger sheets than the granodiorites and overall were far more homogeneous in nature. In the Sruhanavarnis Valley, this phase was abundant within the Glenveagh 3A raft-zone and also very common between this raft-zone and the Derryveagh 2 raft-zone (the old southern margin of Thorr). It may have been these larger tonalitic sheets which were responsible for separating the raft-zones in the central areas of the pluton, (see figure 8:2) i.e. these sheets may have been up to 50-100 metres in width. Again the exact size of these sheets is uncertain due to disruption by the subsequent emplacement of the monzogranites.

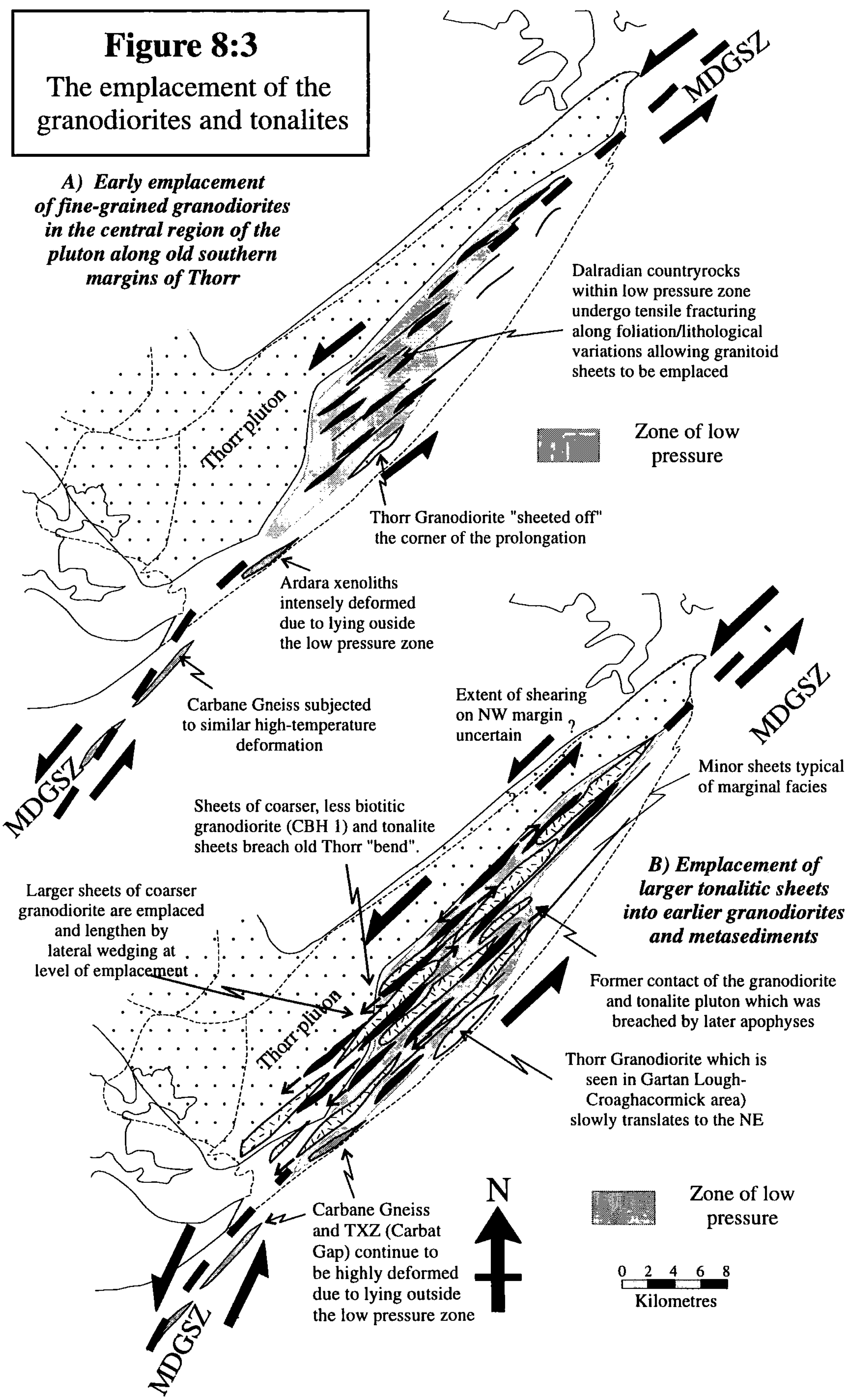
The marginal granitoid facies and the sheeted country rock may have well have formed throughout the emplacement of the granodiorites and the tonalites. The evidence presented in chapter 5 suggests that the monzogranites are younger than the marginal facies as where these porphyritic monzogranites are seen near the margins they never contain the same volume of pegmatitic material as do the marginal granitoids. Furthermore, along the NW margin the transition from marginal facies into porphyritic monzogranites is quite abrupt implying that the latter may have intruded through the marginal granitoids. This might imply the marginal facies was contemporaneous with the granodiorites and tonalites of the central areas.

8:2:3 The emplacement of the early granodiorites and tonalites.

Establishing the true geometry of these early sheets it is very difficult due to disruption caused by intrusion of later porphyritic monzogranites. It is clear, however, that prior to the intrusion of the porphyritic monzogranites, these early tonalites and granodiorites were essentially competent as indicated by the relatively sharp contacts seen in the field. The degree of deformation within these early granodiorites and tonalites is weak to moderate with quartz being the only mineral displaying evidence of crystal plastic deformation. Magmatic evidence for deformation is only slight with only very faint PFC and tiling fabrics. Fracturing of plagioclases is visible in some sections and may be indicative of sub-magmatic fracturing as described by Bouchez *et al.* (1992). The overall lack of deformation in these early granitoids may suggest they were essentially accommodated within the shear zone.

In figure 8:3 the distribution of the fine-grained granodiorites is shown with them essentially occupying the central regions of the present outcrop of the Main Donegal Granite. It is interpreted that a low pressure zone may have developed within

Figure 8:3
The emplacement of the granodiorites and tonalites



this area of the MDGSZ which attracted granodioritic material into to this vicinity where it then intruded along the bedding/foliation within the Dalradian metasediments. The old "bend" in the Thorr contact may correspond to the western extreme of this low pressure zone with rates of country rock dilation decreasing to the SW. This low pressure zone is also believed to have been present to some degree during the intrusion of the tonalites. In chapter 5 it was stated that the Binaniller and Crockmore apophyses may have intruded across a former contact of the granodiorite and tonalite pluton. In the Barnes Gap the granitoids of these apophyses are essentially intruded into metasediments with the Crockmore granite forming the immediate margin to the pluton. At Leahanmore and Croaghacormick the margin of the pluton is composed of banded granitoids and abundant pegmatite which are typical in appearance to the marginal facies of the NW margin of the Main Donegal pluton with the Crockmore monzogranite being encountered 0.75 km away from the pluton contact. Therefore in the intervening ground between Leahanmore the former contact of the earlier pluton must have been breached by these two later apophyses. It was noted that within these septa and raft-trains, which outcrop either within, or at the margins of these two apophyses, the degree of permeation by the host granitoids is only minor which further suggests the absence of the granodiorites and tonalites in this area i.e. the metasediments have not been subjected to such high temperatures related to granitoid emplacement. The heavily permeated nature of the rafts in the central areas my be due to a protracted history of granitoid emplacement in this area and may not be related to these rafts being deeper within the pluton as Pitcher & Read (1959) had originally envisaged.

This possible breached earlier contact of the granodiorite and tonalite trends in a more northerly direction which is broadly parallel to the former "bend" within the Thorr prolongation. These two "bends" may therefore define the eastern and western limits of the low pressure zone which occupied the present central areas of the pluton (see figure 8:3). One must emphasise that this low pressure zone is not a pull-apart as there is abundant evidence of metasedimentary material within the area between these two bends. Instead it is interpreted as zone of dilation which allowed emplacement of granitoid sheets. Within such a low pressure zone the country rock is likely to have undergone tensile fracturing related to tectonic dilation and a combination of magmatic pressure with dilatancy along anisotropies, which in the case of Dalradian metasediments would probably be lithological variations. It is well known that rocks are relatively weak under tensile fractures (≤10 MPa) (Etheridge 1983) and (McNulty *et al.* 1996). These tensile zones would then be easily exploited by invading granitoid sheets and wedged further apart by the hydrostatic pressures of these magmas. In the literature there are many examples of plutons which have been emplaced in low pressure zone i.e. pull-aparts, transfer zones, and releasing bends e.g. The Doctors Flat pluton, Australia (Morand 1992); Pombal pluton, Brazil (Arcanjo *et al.* 1994) and Mortagne pluton (Guinebertau *et al.* 1987).

The Thorr Granodiorite at Gartan Lough may have originally formed part of the southern "corner" (i.e. between Barra Bog and Glenleheen) of the prolongation (see figure 8:4) but subsequent granitoid sheeting during the early construction of the Main Donegal Granite may have led to it becoming sheeted off the main mass. If this was the case then the Falcarragh Limestone and the Sessiagh-Clonmass Fm. which the Thorr material is in contact with must then have been originally situated further to the west than their present location. If this displacement did occur then it would not necessarily disturb the country rock stratigraphy as the direction of opening (ENE-WSW) within this low pressure zone is almost parallel to the strike of the regional stratigraphy. Restoration of this Thorr Granodiorite "fragment" would suggest that 15 km of displacement may have occurred during the development of the low pressure zone. The amount and direction of opening relative to the strike of the regional stratigraphy would be in the region of 1 km, a value that is not unreasonable with regards to the present distribution of the country rock stratigraphy. It must emphasised that the value of 16 km is derived from map evidence and considers displacement after the whole pluton had been constructed and not from the emplacement of the granodiorites and tonalites alone.

During the emplacement of the tonalites and granodiorites in the low pressure zone it is interpreted that these sheets may have laterally expanded at the level of emplacement, along the NE-SW striking anisotropy of the country rock, due to magmatic pressure during the injection of these sheets. It was probably this high magmatic pressure which caused the "bend" in the Thorr prolongation to be breached with high tensile stresses developing at the tip of the sheets which allowed fracturing of the older Thorr pluton to occur. The probable presence of NE-SW trending countryrocks probably provided a well-defined "grain" along which the magmas may have rapidly channelled. This grain in the country rock is at high angles to $\sigma 1$. For granitoids sheets to exploit these planes of weakness the magma pressure must exceed $\sigma 1$ and under these conditions the fractures at the sheet tip could develop in any direction depending upon the most favourable weakness within the host rock, (Fowler (1994). Strongly tensile stresses at the tips of propagating sheets are sufficient to fracture most isotropic rocks oriented normal to the melt-filled plane (Brisbane 1986). Therefore the propagation of the granodiorite and tonalite sheets along the Dalradian metasediments towards the Thorr bend, which is oriented at approximately normal to the foliation in the metasediments, led to it being breached by the invading sheets. The larger width of the tonalite sheets may be related to increased displacements within the shear zone resulting from melt-enhanced deformation or generally higher temperatures within the country-rock allowing more continuous lower strain rate ductile deformation to occur. The process of melt enhanced deformation may lead to the development of a positive feedback situation where the melt within the shear zone allows further displacements to occur along the low-pressure zone. This could then attract more magma into this area which subsequently facilitates even more deformation, and so on. This phenomena implies that once melt enters a shear zone, the process of tectonic space creation may not be as important as factors such as magmatic pressure and "head" (i.e. availability of melt within the source) start to dominate the style of emplacement. The progressive heating of ascent conduits by continued magma flow may allow more granitic magma to ascend without freezing (Petford (1996). This might explain why the earlier granodiorite sheets were quite small although during later intrusions the ambient temperatures from the source to emplacement levels begin to rise allowing the formation of progressively larger sheets (e.g. the tonalites).

With the development of a low-pressure zone the strain must be accommodated at either end where the two splays of the shear zone unite. In the present area of the Main Donegal Granite, this zone would occur between the Glenleheen-Glenties area to the SW and in the Newbridge-Glen area to the NE. In the former, the high temperature deformation within the Carbane Gneiss and the Ardara xenoliths at Carbat Gap may have been caused by further high strains related to the development of the low pressure zone within the MDGSZ. To the NE, the ambient strains are generally much higher, which Hutton (1982) claimed to be the result of greater displacements in this area. Here his strain calculations include the emplacement of the monzogranites, i.e. some of this strain in this area may be related to further displacements during and after the emplacement of the whole Main Donegal pluton.

It is believed that the MDGSZ is a feature mainly of the middle crust (10-13 km). To what extent this feature extends into the lower crust is uncertain. This raises quaestions as to how the granite sheets may have ascended from the source region to the level of emplacement. Hutton (*pers. comm* 1997) states that there may have been some lineamental control on the emplacement of some of the Main Donegal Granite. Features such as the shear-zone culmination of Hutton (1982) (discussed in chapter 3) may be related to smaller lineamental features which are parallel to the main Donegal lineament sited below the five plutons further to the west. Furthermore, the presence of the NNE trending bend in the Thorr prolongation may be related to a feature within the basement. The complimentary eastern bend of the low pressure zone may be associated with a similar lineament which may have exerted subtle controls on the development of the low pressure zone. These smaller features may also have served as conduits to felsic melts and may have ultimately governed the spatial siting of sheets. Figure 8:4 is a schematic block diagram through the crust showing the presence of a series of parallel lineamental conduits in the lower crust which may channel felsic melts. Where these lineaments encounter the Dalradian strata in the middle regions of the crust which lie at a moderate angle to these basement features the architecture of the higher crustal levels will become important in governing the orientation and morphology of such sheets with granitoids exploiting pre-existing anisotropies.

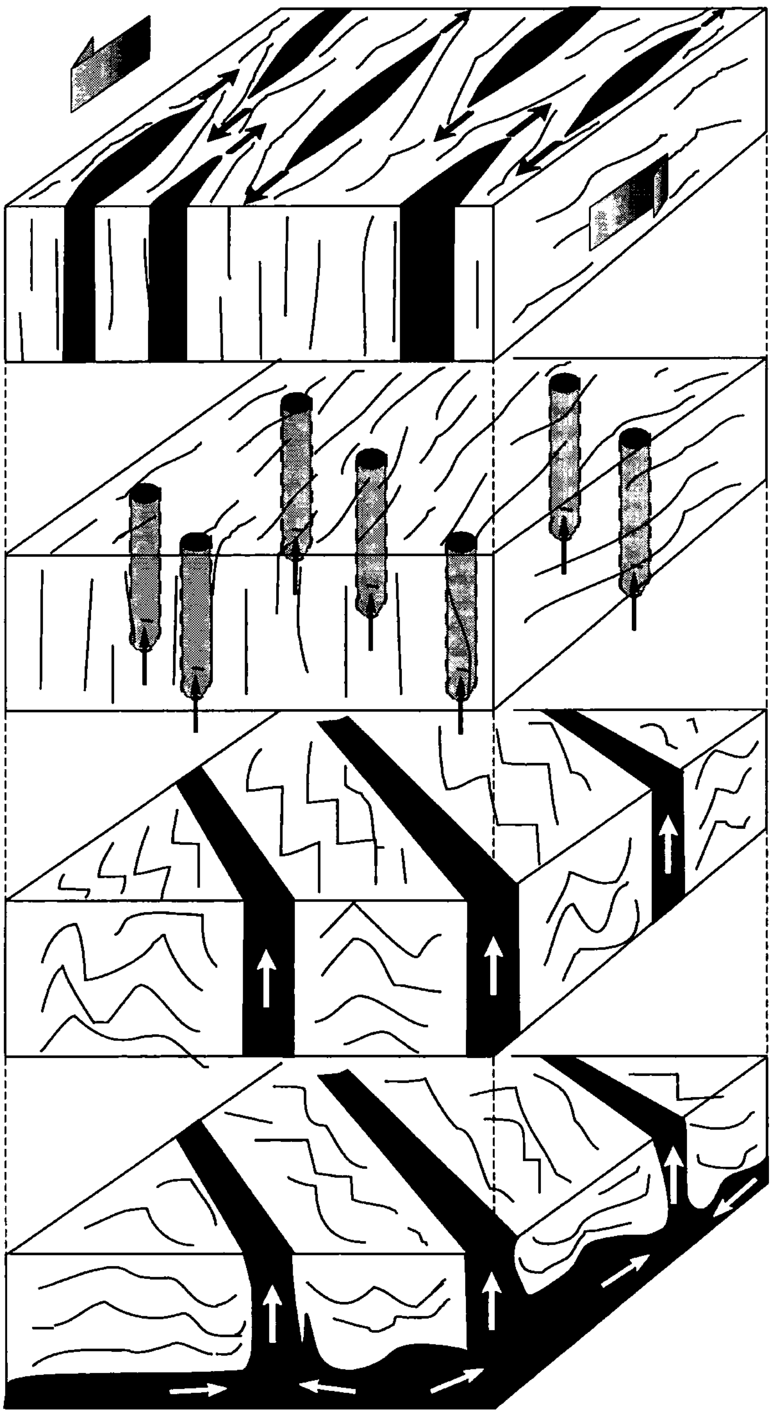
The influence of lower crustal lineaments and higher crustal shear zones in controlling the ascent and emplacement respectively of granitic plutons is gaining support by many workers. In the "Argyll Suite" of granite plutons in the Scottish Highlands Jacques & Reavy (1994) stated that the presence NW-SE trending pre-Caledonian basement lineaments controlled the siting of these plutons at the intersections of more higher level Caledonian, NE-SW trending shear zones. Although this idea was initially suggested by Watson (1984) for the British Caledonian Granites new research is favouring the basement lineament model in the possible siting of plutons, for example, the granite plutons of the "Cairngorm Suite" in East Scotland (Tribe *pers. comm.* 1996).

The marginal granitoids of the Main Donegal Granite

The exact timing of the sheeting in the country rocks is uncertain because these more marginal areas are separated from the central areas by the younger porphyritic monzogranites. However, the author believes that the bulk of these marginal sheets had been emplaced prior to the emplacement of the monzogranites. To what extent the marginal sheeting is related in time to the granodiorites and tonalites is uncertain, i.e. were they intruded synchronously or were they intruded later? There is a strong possibility that the marginal granitoids are the same age as the tonalites in the central areas. Despite the Thorr prolongation having been broken up by the porphyritic monzogranites there is every possibility that the prolongation had been heavily impregnated by the granodiorites and tonalites prior to the emplacement of the monzogranites. Proving the validity of this statement, however, is more difficult.

In the Galwollie Hill-Croaghleconnel area the deformation within the younger monzogranites is very intense implying that the shear zone was sited in this area during the emplacement of these later granitoids (Trawenagh Bay Granites). In chapter 6, however, the earlier tonalite phases (GWH 1) are weakly deformed. This

Figure 8:4
Schematic representation of the influence of early basement features which facilitate the ascent of granitoid magmas and the control exerted by anisotropies in the higher levels of the crust upon the geometry of intruding sheets



Level of emplacement. Granites migrate laterally exploiting "grain" by combination of tectonic space creation and magmatic wedging by fracture propagation

Dalradian Cover with "grain" oriented differently to basement lineaments. Hence ascending granitic magmas exploit weaknesses within the cover i.e. structural grain or lithological variations

Basement with lineaments acting as conduits (felsic dykes)

Basement/Source (granitic melt generation) may suggest that the north-western part of the shear zone may have not deformed the granitoids within this area; it is possible that it was still being "pinned" against the NNE trending lineament. The absence of large volumes of tonalite and granodiorite in the Galwollie Hill-Croaghleconnel area makes the true extent of the MDGSZ in this area difficult to be ascertain during the emplacement of these two earlier phases. Furthermore the involvement of the main NNE Donegal lineament during this early stage in the construction stage of the Main Donegal Granite is also uncertain.

8:2:4 The porphyritic monzogranites

The porphyritic monzogranites form the greatest volume of granitoid material within the Main Donegal pluton, being most abundant in the central and north-western parts of the pluton. Their field relationships between earlier granitoids, i.e. sharp, planar and angular contacts, imply that the early granodiorites and tonalites were relatively competent and capable of brittle fracture during the emplacement of the monzogranites*. This feature also implies that the construction of the Main Donegal Granite may have been somewhat episodic. Furthermore the change in the chemistry of the melts from tonalites to more potassium and silica-rich monzogranites implies some degree of time may have elapsed to account for the change in character of the melts in the source. The outcrop relationships between the different porphyritic monzogranites suggests that they may have been emplaced quite rapidly with no sharp contacts observed anywhere between the respective monzogranites. Instead these contacts tend to be transitional or separated by banding with this latter feature attributed to syn-magmatic deformation in the form of magmatic shears, localising along contacts due to the presence of subtle viscosity contrasts. The varied appearance of the monzogranites may be a feature related to the source, or alternatively may be due to subsequent alteration caused by interaction of the magma and the wallrock during ascent. The relative chronological relationships between the porphyritic monzogranites is uncertain, due to the difficulty in identifying contacts between the respective sheets in the field and hence identifying cross-cutting relationships.

* In Map A the lensoid nature of the medium-grained CBH 1 granodiorite suggests that *some* of these earlier phases may not have been fully competent but, nevertheless, were capable of fracturing due to high magmatic pressures, i.e. CBH 1 shows Bingham properties. When this magmatic pressure had been released (after the catastrophic emplacement of these monzogranites) the rafts deform in a more viscous fashion producing their lensoid shapes.

8:2:4:1 The porphyritic monzogranites of the central zones

The size of the monzogranite sheets in the heterogeneous, central areas appears to range from 50 metres to 400 metres in width and tend to contain abundant rafts of metasediment, granodiorite and tonalite. This common occurrence of rafts of the earlier granitoids implies that the monzogranites were emplaced under high magmatic pressures leading to catastrophic fracturing of these earlier phases. In all three mapping areas it was demonstrated that the rafts of the earlier granitoid were separated both vertically and horizontally; a similar relationship had been suggested for the country rock rafts within the pluton (Pitcher & Read 1959). The intrusion of the monzogranites may therefore have resulted in a reduction in the competency of this early formed pluton causing strain to be localised into the more viscous crystallising monzogranite melts. As a result the distribution of strain became more homogeneous within the pluton, i.e. strain was being accommodated within the cooling monzogranites. As the pluton progressively cooled strain may then have concentrated within the marginal areas of the pluton as is now presently seen. Early country rock, granodiorite and tonalite fragments can be envisaged as rafts floating within a "sea" of porphyritic monzogranites of varying viscosity due to different stages of cooling. The alignment of rafts may result from minor degrees of rotation (i.e. analogous to PFC fabrics alignment) into the shear plane although from the field evidence shown in the Maps A and B, this alignment may be related to fracturing orientations during actual emplacement. The presence of rafts within a viscous monzogranite host may account for the majority of the banding types seen within the pluton.

- *The regular banding at Doocharry*:- this banding is believed to have developed around a core of competent, fine-grained granodiorite. The dark bands are composed of tonalite/trondhjemite whilst the lighter bands are typically porphyritic monzogranites. Berger (1971) claimed that the light bands were formed by the addition of potassium-rich fluids along zones of dilation which caused the tonalites to be converted to monzogranites. In this study it is interpreted that the banding was produced by zones of relatively high magmatic straining within the pluton. The footnote on page 359 describes field evidence that possibly implies some of the tonalites and medium-grained granodiorites were not fully competent during the intrusion of the monzogranites and display Bingham properties i.e. capable of fracture in essentailly a zone of high plastic flow. At Doocharry this visco-plastic behaviour within the tonalite may have occurred. The fine-grained granodiorites seen elsewhere in the pluton were predominantly competent during the intrusion of the tonalites. Therefore, this observation may explain the occurrence of the fine grained mass which forms the core of the Doocharry synform. The surrounding tonalites however may have been capable of fracture under high effective stresses, but viscous flow during lower effective stresses. Therefore after the emplacement of the porphyritic monzogranites the relatively viscous tonalite was probably capable of being smeared out to form the dark bands. Berger (1971) states that the dark bands are rarely more than 3-5 metres in length and suggests that this is evidence for the dark bands being older. Comparison of the tonalites from the Sruhanavarnis Valley and the Glendowan Mountains with the dark, regular bands shows that they are essentially identical. Therefore the regular banding is interpreted to have formed in areas of the pluton which were undergoing relatively high strain magmatic deformation. In the Sruhanavarnis Valley, localised banding is seen around rafts of countryrock and earlier formed granitoids. The width of banding appears to be controlled by the size of the rafts and the magnitude of syn-magmatic deformation.

- The syn-magmatic shears (cross bands of Berger 1967) probably developed later when the porphyritic monzogranites were below their RCMP values. This led to localising of strain into discrete shears which became filled with interstitially derived melt. Within the Main Donegal pluton the orientation of syn-magmatic shears is mostly parallel to the foliation (i.e. 058°) and therefore implies some external tectonic control.

8:2:4:2 The porphyritic monzogranites which form homogeneous zones

The internal nature of the more homogeneous zones, i.e. the pink porphyritic monzogranites of the NW region of the pluton and the Binaniller apophysis, implies that these larger units may well be later and may have been intruded into a slightly higher level within the crust with the "stoping zone" of these larger sheets having been eroded leaving these units relatively free of any other inclusions (see next section). Alternatively these monzogranites may have emplaced themselves by a combination of tectonic space creation and magmatic wedging

The distribution of the monzogranites implies the loci of emplacement changed during the construction of the pluton with the more homogeneous zones interpreted as being zones of large-scale intrusion where large volumes of compositionally similar magma have been emplaced quite rapidly with greater homogenisation occurring at the level of emplacement. This would account for the lack of sharp internal contacts within these homogeneous zones.

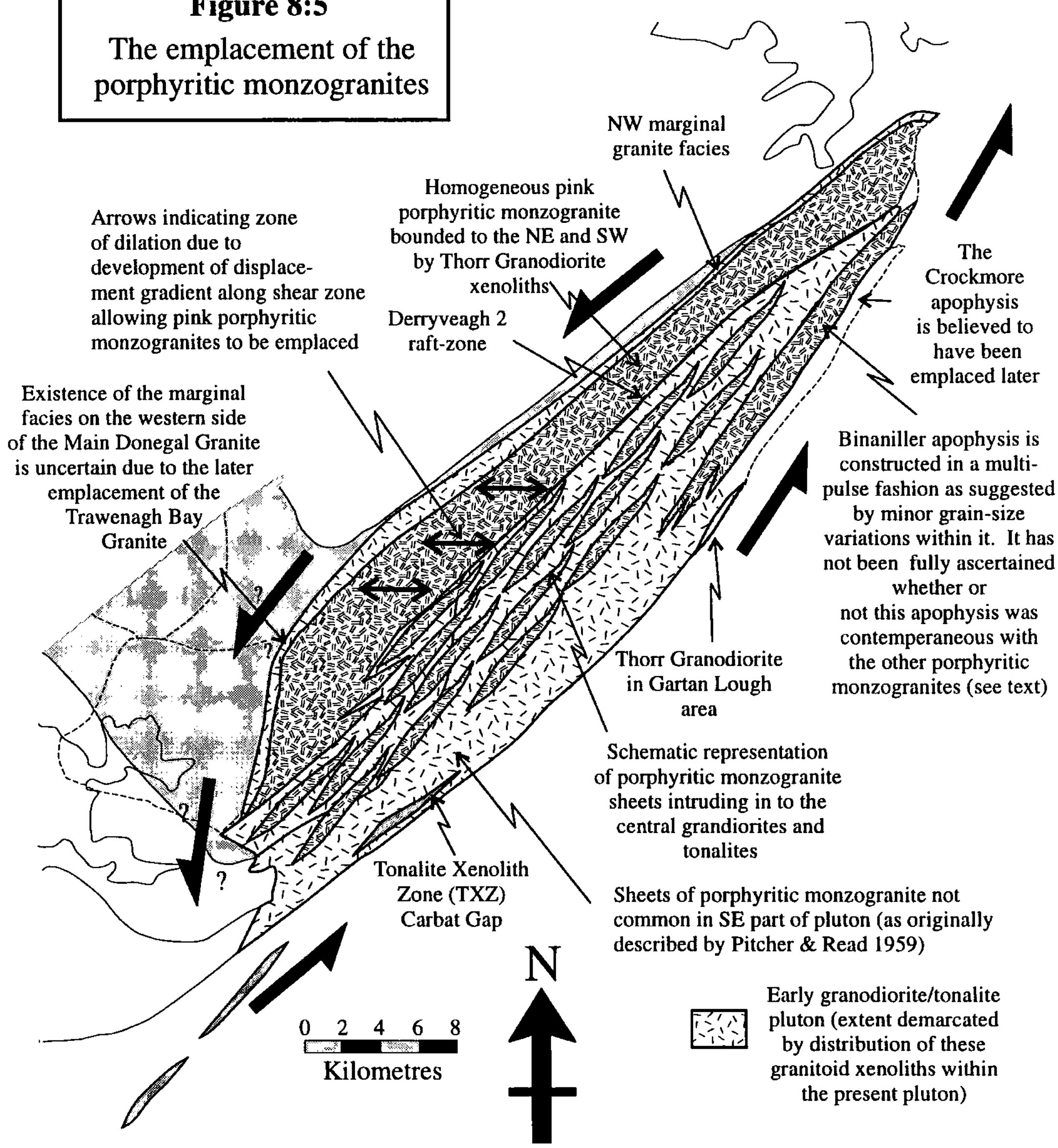
8:2:4:3 The emplacement of the porphyritic monzogranites

The initiation of porphyritic monzogranite emplacement is uncertain. Angular relationships between these and the granodiorites and tonalites implies there may have been a period of relative quiescence along the MDGSZ. Therefore an assesment of what process that caused the MDGSZ to become more operative again are required. Possibilities include the presence of monzogranitic melt within the shear zone which caused it to reactivate, or, alternatively movements along the shear zone (i.e. regional tectonic factors) allowed the monzogranites to be emplaced. The answer is not known and the author cannot recollect any field evidence which may shed light on this problem. It is clear, however, that there is evidence for syn-magmatic deformation during the emplacement of these monzogranites, i.e. banding and syn-magmatic shears.

A combination of high magmatic pressures and abundant melt within certain parts of the shear zone may have led to the development of a displacement gradient along the length of the shear zone as Hutton (1982) had originally envisaged. The emplacement of large volumes of porphyritic monzogranitic sheets in the central areas (see figure 8:5) of the pluton within the low pressure zone (which allowed the emplacement of the granodiorites and tonalites) may have led to the positive feedback situation described in section 8:2:3. i.e. magma within a shear zone facilitate melt-enhanced deformation which then accommodated increased displacements at lower strain rates. The increased displacements resulting from the intrusion of the smaller porphyritic monzogranite sheets in the central areas may have contributed to the development of the displacement gradient. This gradient caused part of the shear zone to split axially (more noticeably the NE side of the pluton) which may then have allowed the emplacement and rapid construction of the more homogeneous masses by continued addition of compositionally similar pulses. Hutton (1982) proposed the NW splay of the shear zone had a more pronounced "bend" due to more outward bending. The large sheet of pink porphyritic monzogranite in the NW part of the pluton may have therefore have started to fill the developing space provided, due to the progressive outward bending of the shear zone. The development of tensile stresses within the pluton related to the displacement gradient would have lead to further exploitation of sheets in the central areas. A combination of tensile stress and magmatic pressure may explain why earlier contacts between different granitoids and country rock have been extensively exploited by later granitoid phases, i.e. contacts are weaker than homogeneous rock due to subtle difference in viscosity and mechanical strength.

In the above model for the development of the displacement gradient it is of importance to determine what was happening in the western extreme of the pluton, (i.e. Brockagh- Galwollie Hill); i.e. was the shear zone active in this area, or was it still being "pinned" against the NNE lineament? If it was being pinned, then by what mechanism were the increased displacements along the MGDSZ being

Figure 8:5
The emplacement of the porphyritic monzogranites



accommodated? For this reason the uncertain relationship of the pink porphyritic monzogranites of the NW margin to the G I pulse of the Trawenagh Bay pluton in the area to the south of the Meenderryherk Septum creates a problem, i.e. is there evidence of the MDGSZ operating in this area during the emplacement of the pink porphyritic monzogranites, but prior to emplacement of G I (Trawenagh Bay Granite)? This area warrants further mapping to obtain the relationship of G I to the pink porphyritic monzogranites as it may allow the movements of the MDGSZ in this area to be constrained (i.e. was it still being pinned or was it affecting the granitoids of the Galwollie Hill prior to the emplacement of the Trawenagh Bay pluton.

The distribution of the porphyritic monzogranites within the Main Donegal Granite indicates that the locus of emplacement was changing. This change may be related to varying displacement gradients along the shear zone with new areas of dilation developing, which are subsequently exploited and filled with granitic magma. The Binaniller apophysis on the south-western margin at the NE end of the pluton is an example of the locus of emplacement shifting. The exact relationship to the other porphyritic monzogranites is uncertain because of subtle differences in petrography (addressed in chapter 7). The relative homogeneity of this apophysis implies that it was emplaced quite rapidly in a similar fashion to the pink porphyritic monzogranites of the NW margin.

The Crockmore apophysis which lies to the SE and close to the margin of the pluton is overall less deformed than the Binaniller apophysis and is believed to be younger in age because the lower strain deformation fabrics present implies it was emplaced at late stages of progressive S_6 deformation. The more aphyric nature of this Crockmore Granite is also a feature not consistent the contemporaneous emplacement of the porphyritic monzogranites. The CBH 3 aphyric monzogranite on Crobane Hill, which post-dates all the porphyritic monzogranites, may have been emplaced at a similar time to the Crockmore apophysis although the former tends to be less biotitic than the latter apophysis.

8:2:5 The cooling of the Main Donegal Granite

During the progressive cooling (below the RCMP) of the porphyritic monzogranites, the mineral alignment would have began to develop within these granitoids. To what extent the earlier granodiorites and tonalites had developed a foliation is uncertain, because the present foliation visible within these earlier phases may have formed during the cooling of the porphyritic monzogranites, (i.e. after the viscosity contrast had diminished between the monzogranites and earlier granitoids). During this cooling of the pluton the strain distribution changes from homogeneously foliated granite to zones of higher strains which lead to the development of S-C' fabrics and in some areas, ultimately to the development of mylonites (Gapais 1989). In the case of the Main Donegal Granite the majority of this deformation was concentrated along the margins of the pluton in response to the granite becoming increasingly competent whilst the foliation within the central MDG was relatively weak. Although mostly weak there are zones of high strain which correspond to sinistral deformation within the pluton itself. These high strain zones within the pluton are commonly seen in the vicinity of countryrock septum/raft-zones, for example the Crockmore septum, the Derryveagh 2 raft-zone (NW end of Sruhanavarnis Valley), the Binaniller 3B raft-zone (SW of Moylenanav) and the Glenveagh 3A raft-zone (Map B: Sruhanavarnis). Evidence from the study of raft-trains within the Sruhanavarnis Valley has shown that they have been subjected to high-temperature deformation (e.g. the metadolerites in figure 4:15 of Chapter 4). Whether this deformation within these rafts relates to movements prior to, or during the emplacement of the Main Donegal Granite is uncertain. Within the pelite rafts there is evidence of down-temperature deformation (i.e. fracturing of quartz and feldspar) which may be the result of strain localising at a late stage within these raft-zones. During the cooling of the pluton these xenolithic zones were acting as solid rafts within the granite with deformation, by viscous flow, of the crystallising granite commonly occurring around them, thus producing the banding. During the continued progressive cooling of the pluton there will have approached a point when it became more competent than the raft-zones, (especially in the case of pelitic rafts), with strain then localising into this rheologically weaker material. Zones of high strain may develop localising sinsitral displacements. These zones may widen to affect the immediately adjacent granitoid. In the Main Donegal pluton this feature was observed within the Crockmore Granite above the Crockmore Septum with strong S-C' fabrics developing; the granitoids which lie above the subcropped Binaniller 3B raft-zone are highly strained over a width of 150-200 metres within the Moyelenanav-Crockskallabagh area. Furthermore at the NW end of the Sruhanavarnis Valley, S-C' fabrics develop within the heterogeneous granitoids to the SE of the Derryvaegh raft-zone and within the pink porphyritic monzogranites to the NW of it.

The overall uniform nature of the foliation within the Main Donegal Granite is attributed to two strain fields within the cooling pluton, (Hutton 1982). Within the countryrock the MDGSZ, $k \approx 1$, transcurrent sinistral shear component was dominant with X sub-horizontal. In the granite itself strain data also implies that the foliation was formed by sinistral shear, although there is a component of flattening ($k \approx 0$). Hutton (1982) attributed this bulk flattening within the granite to the superimpostion of another strain component mainly confined to the granite itself, and the immediately adjacent country rock, with the X direction of this strain ellipsoid being vertical. The superimpostion of the shear zone strain ellipsoid and the granite strain component would have caused bulk flattening within the granite producing the S_6 foliation. The strain component within the granite is believed to have been caused by upward movement of the granite, causing overall vertical extension. This upward movement may have been caused by the buoyancy within the granite causing the shear zone to be "jacked-up" (and producing the NNE-SSW trending shear zone culmination or it was related to the lateral confining pressures of the walls to the pluton). In the NE part of the pluton and along the margins Hutton (1982) believed that the shear zone strain exceeded and overcame the granite-strain component with the granites of this area possessing a strong sub-horizontal stretching lineation. The general absence of this granite-strain component within the aureole (apart from a few 10's of metres from the granite contact) was the feature which Hutton (1982) attributed to general permissive emplacement of the Main Donegal Granite in the MDGSZ. During the cooling of the pluton one would expect this component of strain within the granite to diminish with the sinistral component of strain becoming dominant, making it possible it for localised zones of sinistral shear to develop within the pluton itself (see figure 8:6). (e.g. in the vicinity of the raft-zones). S-C fabrics within the pluton tend to be localised in the marginal areas and within localised zones of sinistral shear within the pluton. The absence of S-C' fabrics in much of the central areas may be due to the pluton becoming too competent to be deformed in a homogeneous fashion with deformation becoming heterogeneous and localising within discrete zones. This feature is commonly seen within other syn-tectonic granite plutons (Gapais & Barbarin 1986; Gapais 1989). Displacement within the relatively rigid pluton was also being accomodated along planar zones in which late microgranites dykes and felsites have been intruded. The occurrence of sigmoidal foliations implies some of these intrusions were syn-deformational. At this late stage in the evolution of the pluton, the almost fully crystallised granites can be envisgaed as a series of rigid blocks which were being rotated relative to one another. These late stage microgranitoids would be accomodating this deformation between the blocks. Sinistral displacements along these dykes are generally more common although dextral offsets are also seen, and typically have an approximate E-W trend. These dextral dykes probably correspond to zones of anti-reidel shearing related to the MDGSZ.

During this late stage of marginal deformation within the Main Donegal Granite, the Trawenagh Bay Granite was probably intruding into it further to the west. The abundance of relatively weakly deformed composite xenoliths at Galwollie Hill may imply that much of the MDG, at least in this area, was highly competent.

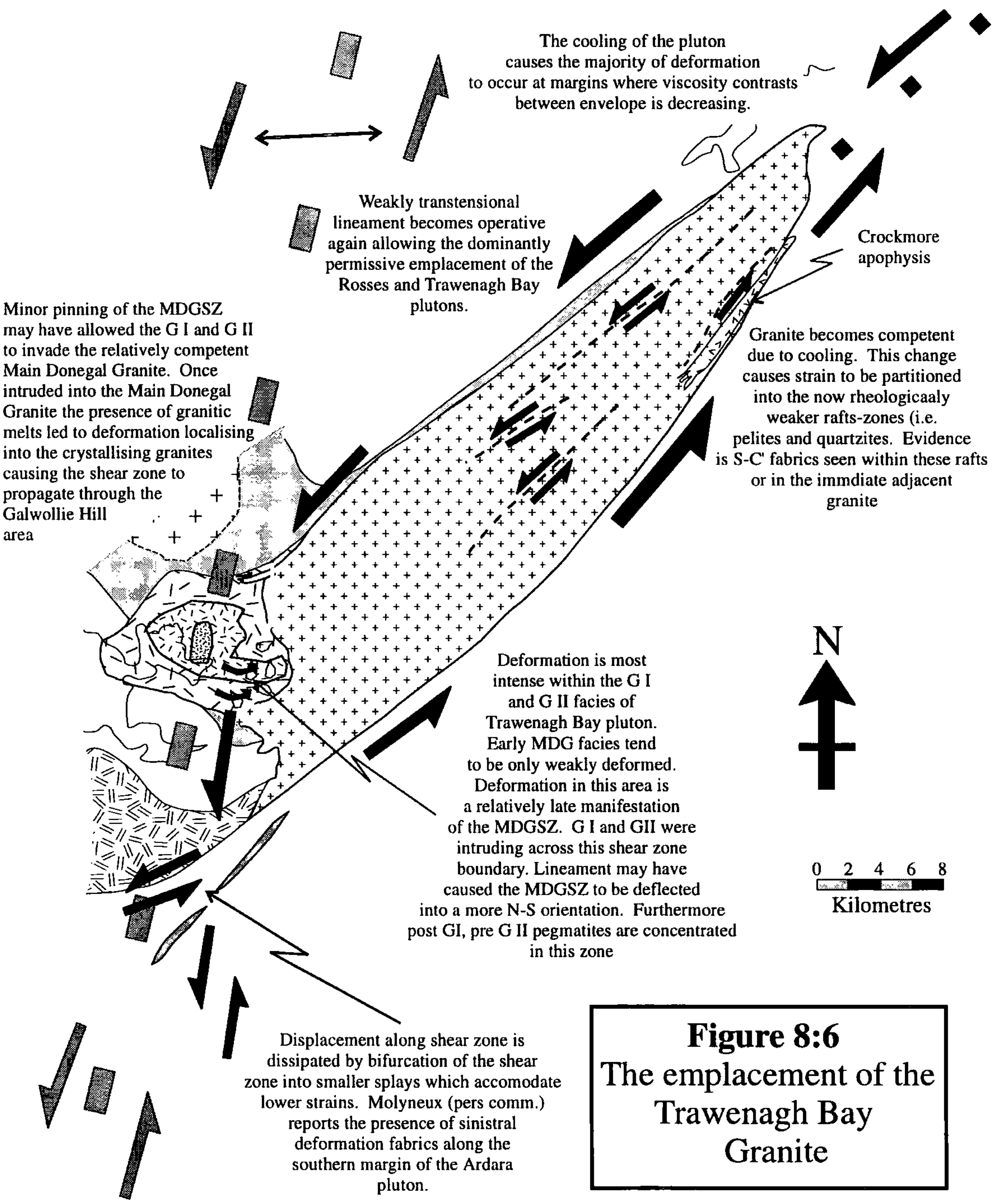


Figure 8:6
The emplacement of the Trawenagh Bay Granite

8:2:6 The emplacement of the Trawenagh Bay Granite

In chapter 3 it was stated that the Trawenagh Bay Granite was essentially passively emplaced, as along the northern, western and southern contacts structural and metamorphic features are only very minor. In chapter 6 the evidence put forward suggests that the Trawenagh Bay Granite is dominantly younger than the MDG although a few crucial relationships remain unresolved (e.g. the relationship of the pink porphyritic monzogranites of the NW part of the MDG and the G I pulse of the Trawenagh Bay Granite). The passive emplacement of this pluton is interpreted as being the result of further movements along the transtensional lineament within a low pressure zone or pull-apart. Similar movements along this lineament may have allowed the slightly earlier Rosses centred complex to be passively emplaced (see figure 8:6). During the initial intrusion of the Trawenagh Bay Granite the north-western splay of the MDGSZ may have still been "pinned" to some degree against the lineament. This may well have subjected the western part of the Main Donegal Granite to tensile stresses and fracturing which allowed the emplacement of the G I pulse of the Trawenagh Bay Granite. The evidence in the Galwollie Hill area suggests that the NW splay of the MDGSZ was also active because the G I and especially G II monzogranites are highly deformed. The presence of granitic melts in this area may have allowed the shear zone to propagate into this area with deformation being accommodated within these viscous melts. The absence of deformation within earlier xenoliths implies that strain was being localised into the later monzogranites (i.e. G I and G II) as they crystallised, and this may have continued into the solid-state. If this MDGSZ deformation occurred in this area after the both the Main Donegal and the Trawenagh Bay plutons had cooled, then one would expect all of the granitoid phases to be deformed equally as viscosity contrasts between the various granites would be minimal. For this reason the eastern part of the Trawenagh Bay pluton is believed to be syn-kinematic in relation to the MDGSZ. The angular relationships and the composite nature of the xenoliths implies that the Main Donegal Granite in this area was totally competent during the intrusion of the Trawenagh Bay Granite. In no other part of the Main Donegal Granite have such relationships been observed. The G I member of the Trawenagh Bay pluton is for the most part emplaced within the Thorr pluton, although in the south-western portion of the pluton the immediate envelope is comprised of Dalradian metasediment. It is possible that this member may have been emplaced by some form of cauldron subsidence mechanism. In the north-eastern part of the pluton a sheet-like mass of G I intrudes the Main Donegal Granite, with it causing a piece of the Thorr pluton to split off to produce the Meenderryherk Septum. The presence of melt in this area may have caused the MDGSZ to then cut across this area with "pinning" along the NNE lineament no longer occurring.

The deflection of the MDGSZ in this area from its usual trend of NE-SW to a more NNE-SSW direction may be related to the presence of the lineament. The influence of the lineament and the MDGSZ may have led to the siting of large volumes of pegmatite in the Galwollie Hill-Croaghelconnenl area subsequent to the emplacement of G I but prior to G II. It is uncertain whether or not the voluminous pegmatites, to the south of this area, in the Derkbeg-Straboy area are related to this same phase observed in the Galwollie Hill-Croaghleconnel area. If they were of the same age then it might suggest that the pegmatites are not necessarily related to the Trawenagh Bay Granite but were intruded during their emplacement due to some structural control (i.e. lineament or MDGSZ) which allowed their ascent and emplacement during the construction of this pluton. The G II phase of the Trawenagh Bay Granite has also been emplaced along the western boundary of the MDGSZ with this phase being more deformed than all of the other granitoid phases present, as seen from the field relationships on Galwollie Hill. The sharp and often angular relationship seen in Map C implies G I and the pegmatites were competent during the intrusion of G II. The highly deformed nature of G II in the Galwollie Hill area implies that the MDGSZ was still very much in operation during the bulk of the emplacement of the Trawenagh Bay Granite. The relationship of the G III pulse (mapped by W.S. Pitcher) which lies within GII in the central area of the presently exposed Trawenagh Bay pluton to the MDGSZ is uncertain as it lies further away from the shear zone boundary. The relatively sharp and planar contacts of G II and G III implies that the former may well have been quite competent during the emplacement of the latter

8:2:7 Summary of the emplacement of the Main Donegal and Trawenagh Bay granites

- Parts of the Ardara and Thorr plutons were emplaced within the Main Donegal Granite Shear Zone (MDGSZ) prior to the emplacement of the Main Donegal Granite itself.
- The emplacement of the entire Thorr pluton implies that both the Donegal lineament and the MDGSZ may have been active at the same time.
- During the early emplacement of the Main Donegal Granite, a low-pressure zone developed within the present central areas of this pluton. The boundaries of the low-pressure zone may have been governed by smaller, parallel, basement lineaments, e.g. the NNE trending bend along the southern margin of the old Thorr prolongation.
- The development of this low pressure zone allowed the emplacement of relatively small sheets of granodiorite along bedding-foliation surfaces within the Dalradian metasediments. Some degree of lateral wedging by fracture propagation at the

level of emplacement may have occurred. The cooling of these early sheets may have been rapid due to intrusion into country rock as indicated by the relatively fine-grain size. Outside the low-pressure zone strains were higher with the Ardara prolongation and the Carbane Gneiss being subjected to the high-temperature solid-state deformation.

- Initial emplacement of the tonalites may also have been controlled by the central low-pressure zone, but greater magma pressures allowed more rapid propagation of the sheets by tensile fracturing along the vertical Dalradian strata. Possible lateral migration of these sheets led to the breaching of the original boundaries to the low-pressure zone.
- Cooling of the granodiorite and tonalite to a competent mass may have led to MDGSZ becoming less operative. Field evidence suggest that some of this early granitoid material may have not been totally competent.
- Change in the composition of the source regions (possible caused by some fractionation process) with monzogranites (mostly porphyritic) becoming the dominant phase to be emplaced.
- Possible further displacements along the low-pressure zone within the central regions of the pluton led to emplacement of relatively small monzogranite sheets, which were under high pressure causing catastrophic fracturing of the earlier granitoids and country rocks leaving them as rafts within the invading monzogranites.
- The presence of monzogranitic melts along part of the MDGSZ may have led to the development of a displacement gradient which caused the shear zone to split in an axial manner. The positive feedback from this displacement gradient may have allowed the rapid construction of the more homogeneous zones of porphyritic monzogranite e.g. the pink porphyritic monzogranites and the Binaniller apophysis. The distribution of the homogeneous zones implies that the locus of emplacement within the pluton was switching.
- Later displacements may have allowed further aphyric monzogranites to be emplaced e.g. Crockmore apophysis and the CBH 3 monzogranite of Crobane Hill. The intrusion of these phases may have been into highly viscous, but not quite competent porphyritic monzogranites.
- The Main Donegal pluton cooled with the NE-SW striking foliation developing. Continual displacements along the MDGSZ now occurred essentially along the margins although there is evidence of smaller zones of sinistral shear within the pluton itself.
- To what extent the Donegal lineament was active during the emplacement of the monzogranites is uncertain.

- During the emplacement of the Trawenagh Bay pluton and possibly the Rosses pluton the Donegal lineament may have been in operation again allowing the predominately passive emplacement of these plutons.
- The eastern part of the Trawenagh Bay Granite was intruding into a competent Main Donegal Granite with angular xenoliths of this latter pluton found in both G I and G II.
- In the vicinity of the MDGSZ these granitoid phases (G I and G II) are intensely deformed, most notably G II. The absence of high strain fabrics within earlier MDG xenoliths implies that strain was being partitioned into the cooling G II magma (i.e. the eastern part of the Trawenagh Bay Granite was syn-kinematic in relation to the MDGSZ). This localisation of strain within the G II phase may have continued into the solid-state with the development of zones of S-C' mylonites.
- The absence of cataclastic textures within the G II pulse and the marginal granitoids of the MDG implies the movement along the MDGSZ may have ceased soon after, or possibly strains were accommodated in shear zones within the aureole e.g. the medial shear zones of Hutton & Alsop (1995) and the DMG_3 shear zones of Pitcher & Berger (1972).

8:3 Discussion

The Main Donegal Granite has for many years been cited as a classic example of a sheeted pluton. This is mainly due to the distribution of raft-zones within the pluton which were interpreted as dismembered roof septa (Pitcher & Berger 1972) and Hutton (1982*b*). Over the past fifteen years or so, the number of documented cases of sheeted plutons has slowly increased; The Great Tonalite Sill, (Ingram & Hutton 1994), Ox Mountains Granodiorite (McCaffrey 1992); the Jackass Lakes pluton, California (McNulty *et al.* 1996) to name just a few. Furthermore it has been shown that sheeted plutons occur within all three of the main tectonic regimes i.e. compression, extension and transcurrent shearing. The sheeted plutons in compressional regimes are generally "space denying" and illustrates that tectonic space creation is not an essential factor which governs the siting and level of emplacement of granite plutons, with magmatic pressure being fully capable of exceeding $\sigma 1$. Granite plutons which have been intruded into transcurrent shear zones (including transpression and transtension) generally tend to be more common. The Main Donegal pluton has also being cited as a classic example of a sheeted pluton which has been emplaced within an active shear zone. Many authors working in similar tectonic settings have made reference to this pluton (e.g. Paterson & Tobisch (1992), Ferré *et al.* (1995), Reavy (1989). The list is long. Recent work has suggested that sheeted plutons may give insights into ascent mechanism with there being an increasing belief that felsic magmas can rise through the crust at a very fast rate in the form of felsic dykes (Clemens & Mawer 1992; Petford 1996). These authors state that plutons can be constructed in a an episodic fashion on time scales as low as 10^4 years by dyke-like conduits. Such dykes may exploit and ascend tectonic features such as basement lineaments or crustal shear zones. Within sheeted plutons one might expect to see sheets which may represent arrested ascent. These aspects will now be discussed.

Despite this long list of sheeted plutons within transcurrent shear zones there have been no cited examples, of which the author is aware of, describing the detailed morphologies of granitoid sheets based on detailed mapping. This present study may be the first of its type. The sheeted nature of the Main Donegal Granite, as many researchers had previously envisaged (Pitcher & Read 1959; Pitcher & Berger 1972 and Hutton 1982), has shown to be essentially correct. This mapping has shown that the Main Donegal pluton displays multitudinous, varied sheets, some of which are easy to identify in the field, whilst other are very subtle.

Throughout the pluton there are zones characterised by specific styles of sheeting. For example, in the central areas of the pluton the granitoids are very heterogeneous, whilst nearer to the margins occur zones which are typified by their overall homogeneity. This study has demonstrated that the central parts of the syn-tectonic Main Donegal Granite possess the most sheeted and complex cross-cutting relationships. It is very difficult to elucidate the mechanisms of emplacement for the earlier units because later movements along the shear zone facilitate further emplacement of granitic pulses which tend to disrupt the earlier formed pulses.

To allow greater understanding of the ascent geometries of these earlier sheets, examination is required of those areas of the pluton where the effects of later sheeting is at a minimum. In the present study these areas appear to be the marginal zones of the pluton where original sheet-like geometries are preserved and thus may provide some evidence of ascent. These small sheets, now preserved only in the marginal areas, may originally have occupied all of the present area of the Main Donegal Granite with some of the earlier xenolithic facies within the central areas of the pluton being the disrupted remnants of these sheets. Therefore evidence of ascent mechanisms within much of the Main Donegal pluton has been destroyed by later granitoid pulses, or syn-plutonic deformation, at the level of emplacement.

It has become apparent that the presence of melt within a shear zone itself promotes the development of a self-perpetuating situation whereby melts localise deformation allowing more rapid and greater magnitude displacements which in turn will allow further volumes of granitic material to be intruded into the shear zone. This process is only arrested by the draining of melts from the source region (Petford 1996). This may result in displacement rates decreasing along the length of the shear zone as the intruded granitoids begin to cool and crystallise. The self-perpetuating situation may resume when sufficient melt has been generated in the source to allow dyke propagation to occur with transport of more granitic material to the level of emplacement. This process is hence cyclic with plutons being constructed in an episodic fashion with the rate of pluton construction ultimately related to the speed at which melts can be generated within the source regions. Mapping in this study has shown that the Main Donegal Granite has been constructed during three episodes of intrusion (four phases including the volumetrically small aphyric monzogranite sheets) with subsequent batches tending to become more voluminous in nature than the earlier ones. The smaller earlier batches (i.e. the granodiorites) appear to be the result of dykes intruding into relatively cold countryrock with crystallisation occurring quite rapidly (as indicated by fine-grain sizes). The relatively rapid crystallisation of these sheets may have been sufficiently quick to prevent the positive feedback of melt enhanced deformation and increased displacements from occurring and hence the volume of this earlier phase was not great. During the second phase of MDG construction (i.e. the tonalites) the crystallisation of these melts appears to be slower. This may be related to the intrusion of the earlier granodiorites having raised the ambient temperature of the host countryrocks. This may have allowed the positive feedback situation of melt-enhanced deformation to occur with larger volumes being emplaced as displacement rates increase allowing the relatively rapid emplacement of larger more homogeneous tonalite sheets. Within the tonalites and granodiorites (excluding the later monzogranites) there is no strong evidence for strong syn-magmatic deformation apart from weak PFC fabrics. No macro-scale banding or syn-magmatic shears were observed. This might imply that displacements along the shear zone was not that large. The field evidence suggests there was a period of relative quiescence with the majority of the tonalites cooling to a competent mass prior to the emplacement. The change in composition from tonalites to porphyritic monzogranites also implies a considerable time-lapse may have occurred between the intrusion of these phases. During the cooling of the earlier granodioritic and tonalitic Main Donegal Granite considerable volumes of monzogranite were being generated in the source region. The initiation of the third phase, i.e. monzogranite emplacement, is believed to have been a catastrophic event with the relatively small monzogranite sheets being emplaced into the central regions. The angular field relationships of the earlier granitoid phases within the monzogranites supports this hypothesis. The intrusion of these monzogranite sheets caused the melt-enhanced deformation positive feedback mechanism to develop with large displacement gradients developing along the shear zone allowing the rapid construction of the large homogeneous monzogranite zones, i.e. the pink porphyritic monzogranites in the NW region and possibly the Binaniller apophysis on the SW margin of the pluton. Evidence of large displacements within the pluton during the emplacement of the monzogranites is indicated by ample evidence of macroscopic syn-magmatic deformation, i.e. regular banding and later the development of syn-magmatic shears as the pluton became more competent with zones of syn-magmatic deformation becoming localised into discrete planar zones.

8:4 Summary

The discussion in the previous section suggests the morphology of granitoid sheets is governed by the *rate of emplacement* which is ultimately related to the rate of magma generation within the source region. The following section will discuss varying hypotheses for differing rates of emplacement associated with features that one might observe. These features are based on observations within the Main Donegal Granite.

- ***Rapid emplacement rates***

  During rapid emplacement of granitic material, the viscosity contrast between respective pulses may be too small to prevent homogenisation occurring at the level of emplacement. Such mixing may destroy evidence of all sheet contacts and also the ascent mechanism. During rapid emplacement it is expected that the composition of the individual pulses will also quite similar due to magmas not having time to evolve. Therefore petrographic features would also be very subtle. This suggests that rapidly emplaced pulses will appear very homogeneous with contacts generally rare. In the present study it is interpreted that the pink porphyritic monzogranites of the NW margin were formed from relatively rapid intrusion. The Binaniller apophysis may also be a similar example although its smaller size may have led to more rapid chilling of earlier pulses which allows grain-size variations to be identified in the field.

- ***Medium to slow emplacement rates***

  At slower emplacement rates the viscosity of earlier emplaced granitoids may be sufficiently different (i.e. more viscous) to prevent mixing and homogenisation from occurring, except along immediate contacts. Therefore the boundaries between these sheets may be highly transitional in nature although subtle variation based on petrographic and textural features may allow different sheets to be identified. This style of sheeting has been identified within the porphyritic monzogranites of the central areas of the pluton. It was noted that these sheets possessed very subtle differences in appearance. The contacts tended to be transitional in nature and commonly represent zones of syn-plutonic deformation due to the presence of minor viscosity contrasts between different sheets.

- ***Slow to episodic emplacement***

During very slow or episodic emplacement earlier sheets may have fully crystallised and become competent and consequently may be capable of being fractured. The presence of abundant granitoid rafts within later sheets may result in major sheet contacts being difficult to find within the field. Only detailed mapping will elucidate the true extent of such granitoid sheets. This feature characterises the relationship of the granodiorites and tonalites with the more voluminous porphyritic monzogranites in the central areas of the pluton. During episodic intrusion, earlier granitic sheets become highly disrupted by subsequent phases and original geometries are often extremely difficult to determine. Also the abundance of earlier granitoid rafts within later sheets makes identification of sheet boundaries very difficult due to the high abundance of xenolithic material.

Therefore the rate of emplacement is clearly going to govern the appearance of sheeted relationships at the level of emplacement and to what extent the earlier phases will have cooled. Due to emplacement rates being time dependent the factor which governs the morphology of sheets is therefore rheology (notably viscosity) of the melt and probably more important the rheology of the host into which the granitoid sheets are intruding into. Temperature has a huge effect on viscosity with it controlling crystallisation rates i.e. the proportion of melt to crystal phases present. Therefore an appreciation of granitoid rheology can help to understand the morphology of sheets which will develop within granite plutons. Other factors which govern rheology are:- strain rate, lithostatic normal stresses (σ_N), fluid pressure (σ) (effective stress σ_N-σ). Combination of these factors can strongly affect the behaviour of a material, i.e. does the host material to intruding granitoid behave in a brittle or ductile fashion in response to these factors.

8:5 Conclusions

- The Main Donegal Granite has been intruded into a sinistral shear zone which is interpreted as being active prior to its emplacement and also subsequent to its final crystallisation. This is based on the study of the adjacent Thorr, Ardara and Trawenagh Bay plutons which have been deformed to varying degrees by the Main Donegal Granite Shear Zone.

- The Main Donegal Granite is a multi-pulse sheeted pluton composed of up to four main granite types which represent episodic emplacement into an active sinistral shear zone :-

    1) Fine-grained equigranular granodiorites. **(Oldest)**
    2) Medium-grained equigranular tonalites.
    3) Medium to coarse-grained porphyritic monzogranites.
    4) Medium-grained equigranular monzogranites. **(youngest)**

- Geochemical analysis of these major phases reveals the oldest granitoid phase to be the most primitive with subsequent phases becoming progressively more evolved.

- The earlier granodiorite and tonalite phases which were originally believed to be sheeted and are now only preserved within the central regions of the pluton as disrupted xenolithic rafts within younger monzogranites. These earlier phases were intruded into Dalradian strata within a developing low-pressure zone developed along the old southern margin of the Thorr Pluton "prolongation".

- This earlier pluton has been extensively disrupted by later porphyritic monzogranites. The self-perpetuating nature of melt within a shear zone leads to the development of increased displacements, subsequently allowing more melt to be intruded into the shear zone with the development of a displacement gradient along the length of the shear zone allowing the emplacement of the large homogeneous monzogranite zones.

- These more homogeneous monzogranite zones which are generally located more towards the margin are interpreted as zones of relatively rapid emplacement with smaller pulses of compositionally similar magma having been emplaced rapidly enough to allow a large degree of homogenisation to occur at the level of emplacement and hence obliterating original pulse contacts. Extensive syn-magmatic deformation occurred within the pluton at this time

- The form of sheeted plutons is governed by the rate of emplacement (the availability and rate of generation of granitic material within the source region to the pluton). An understanding of granitoid rheology of both the intruding magma, and the host into which it is intruding, is important as this governs the appearance of sheets within outcrop.

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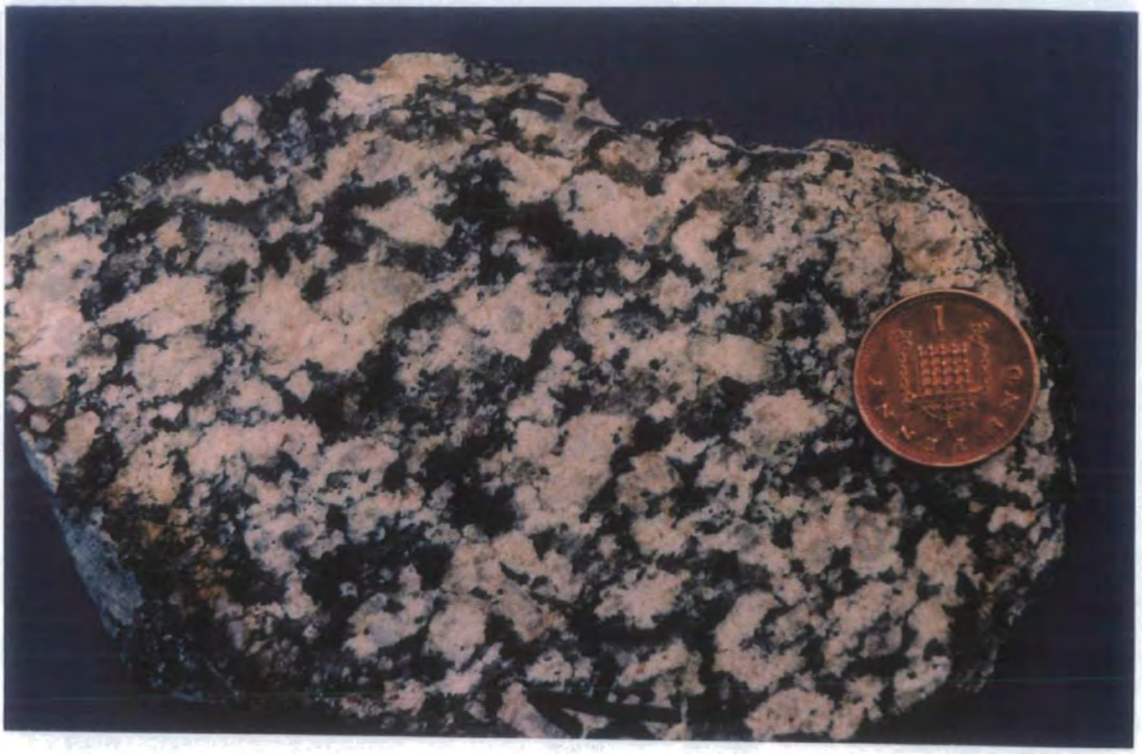
Appendix A

Polished hand specimens of granitoid facies within the Main Donegal Granite

The samples in this appendix have already been referred to within the text of the main part of this thesis. The following photographs illustrate the main variations observed within the granitoid facies from the various mapping areas within the pluton.

The photographs have been arranged in the order in which they were discussed in the thesis.

A) The Granitoids of Crobane Hill (Map A)

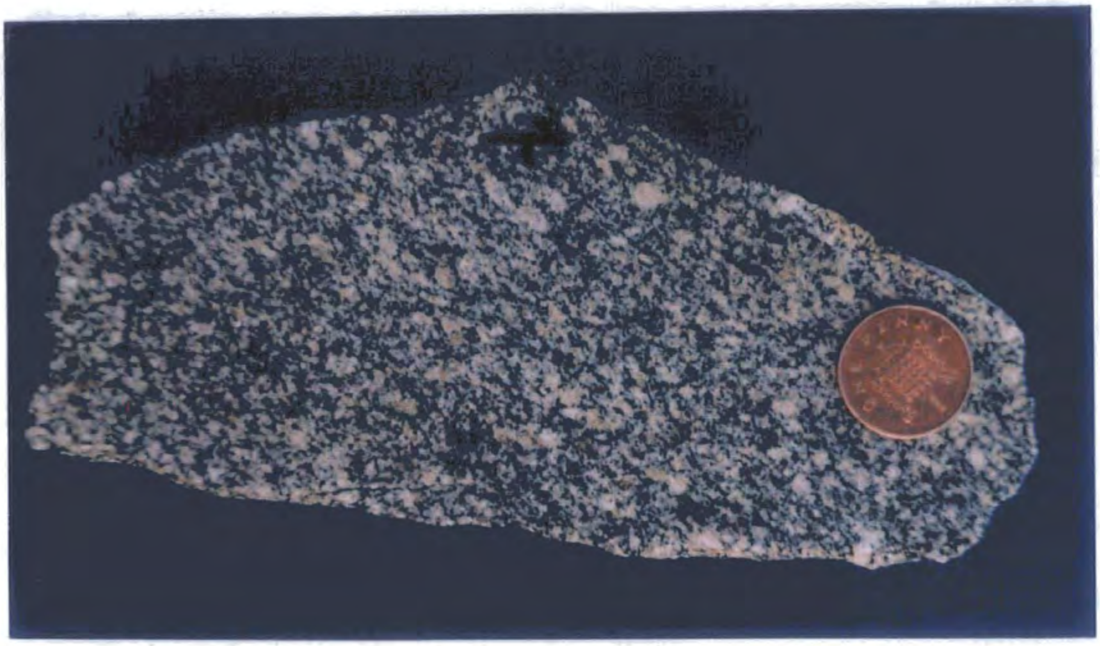


1) Thorr Granodiorite.

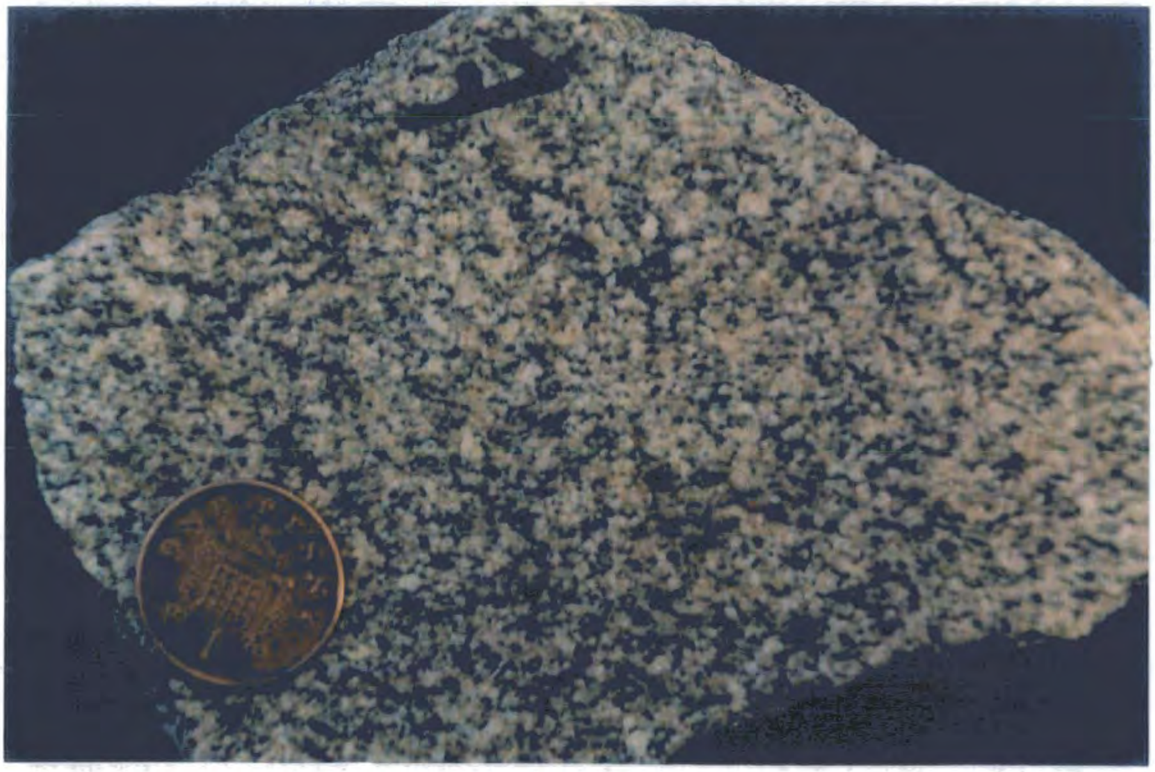
2) CBH 1:- medium-grained equigranular granodiorite.

3) CBH 2:- medium to coarse-grained porphyritic monzogranite.

4) CBH 3:- medium-grained, biotite-poor equigranular monzogranite.



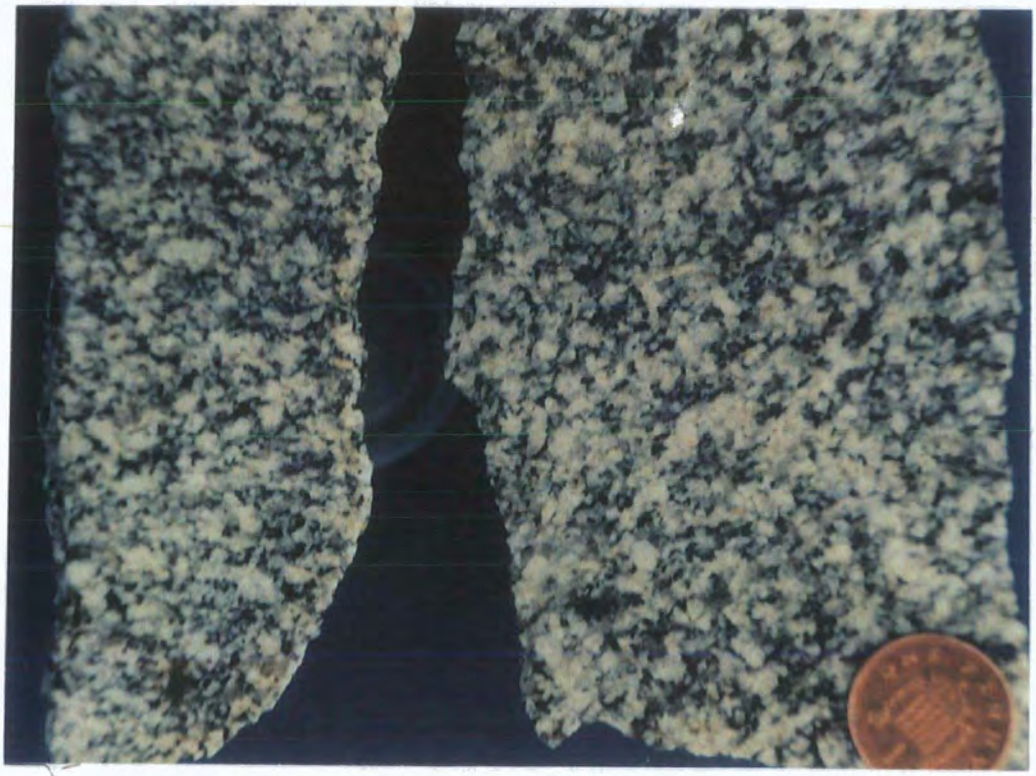
5) **CBH 4/ CBH 4***:- fine-grained equigranular tonalite.



6) **CBH 5**:- fine-grained equigranular monzogranite.

B) The Granitoids of the Sruhanavarnis Valley (Map B)

1) **SRU 1**:- fine-grained equigranular granodiorite.



2) **SRU 2**:- medium-grained equigranular tonalite.

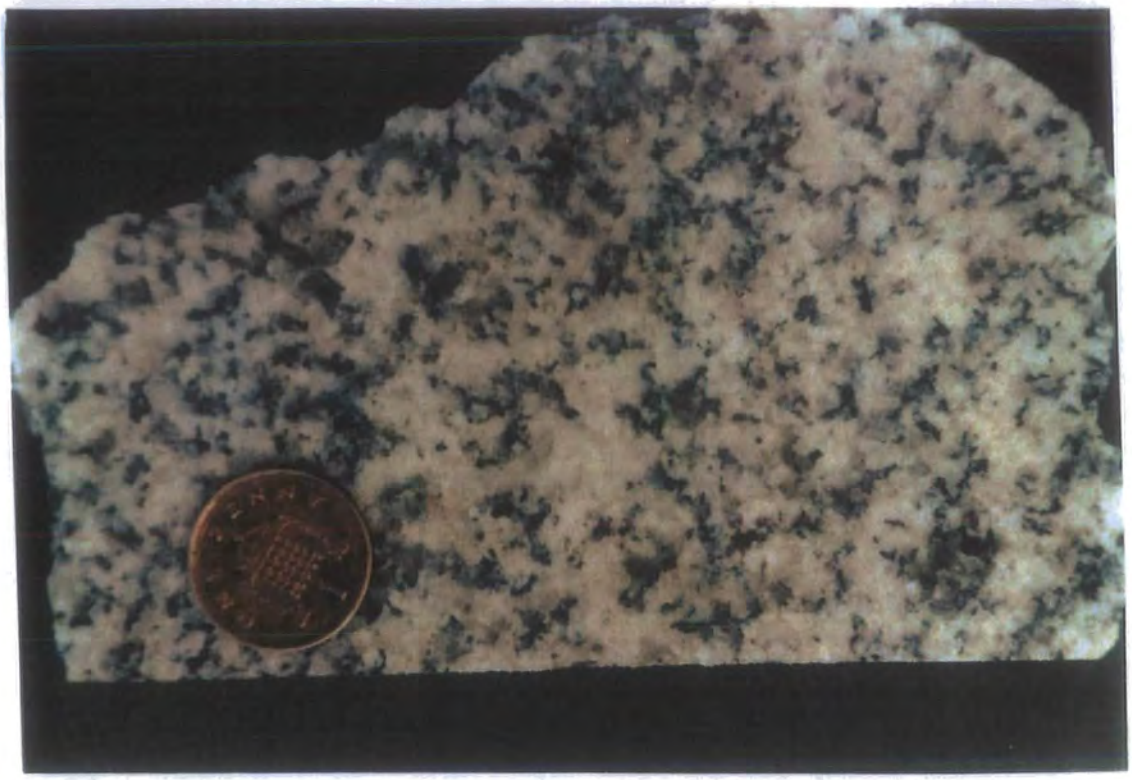
3) The SRU 3 porphyritic monzogranites

i) **SRU 3: PSBG**:- medium-grained porphyritic monzogranite.

ii) **SRU 3; PSBGp**:- similar to i) but contains pink megacrysts of K-feldspar.

iii) **SRU 3: PPBG**:- medium-grained pink porphyritic monzogranite (coarser groundmass than PSBGp).

iv) SRU 3: VCPG:- coarse-grained porphyritic monzogranite.

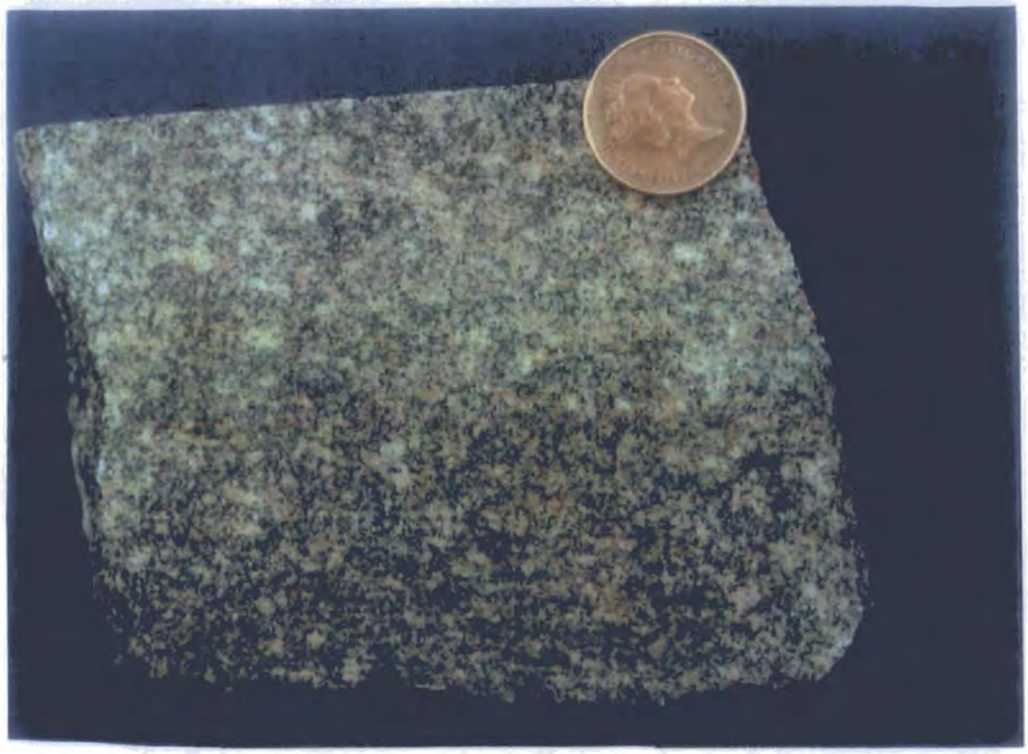


v) **SRU 3: CWBG**:- coarse-grained weakly porphyritic monzogranite.

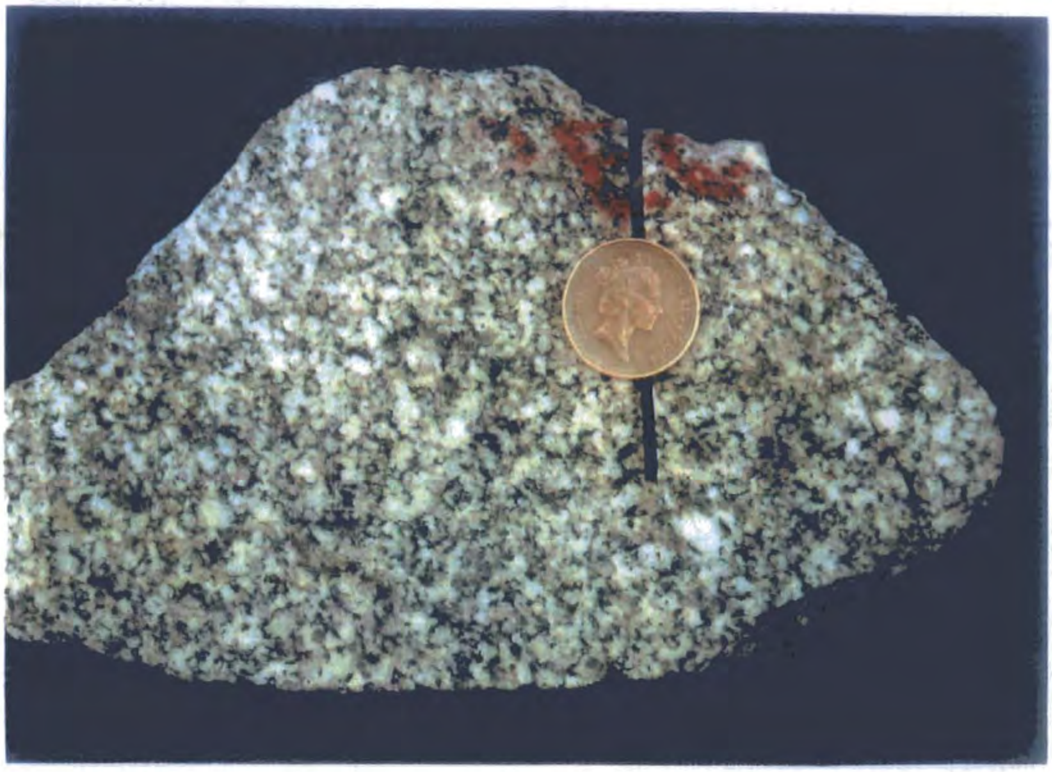
vi) **Comparison of the porphyritic monzogranites of the Sruhanavarnis Valley**. *Top left*; VCPG, *Top right*; PSBG; *Bottom left*; CWBG, *Bottom middle*; PPBG *and bottom right*; PSBGp.

4) Later microgranitoid dykes within the Sruhanavarnis Valley.

C) The Granitoids of the Glendowan Mountains (Maps I and II)



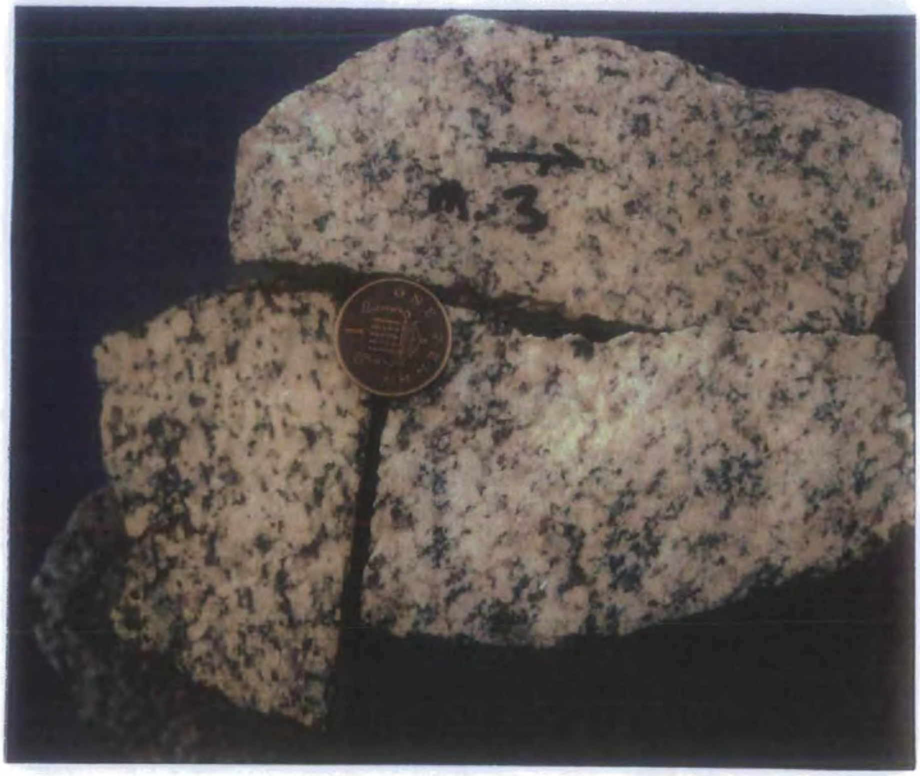
1) **GM 1**:- fine-grained equigranular granodiorite.



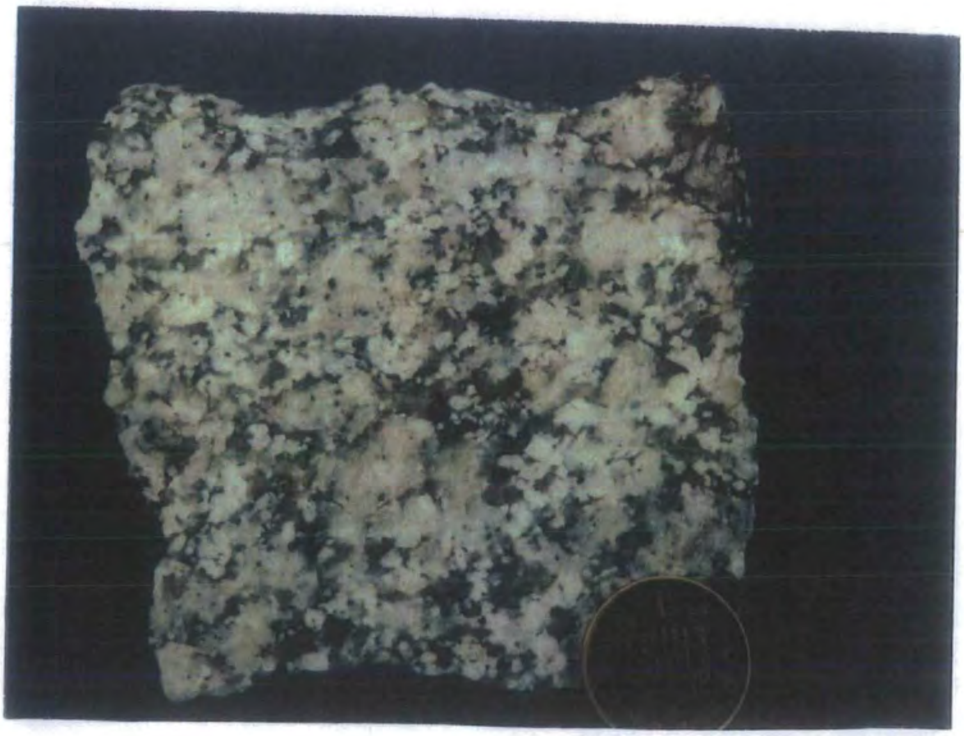
2) **GM 2**:- medium-grained equigranular tonalite.

3) **GM 3**:- fine to medium-grained equigranular monzogranite.

4) **GM 4**: medium-grained subtly porphyritic monzogranite.



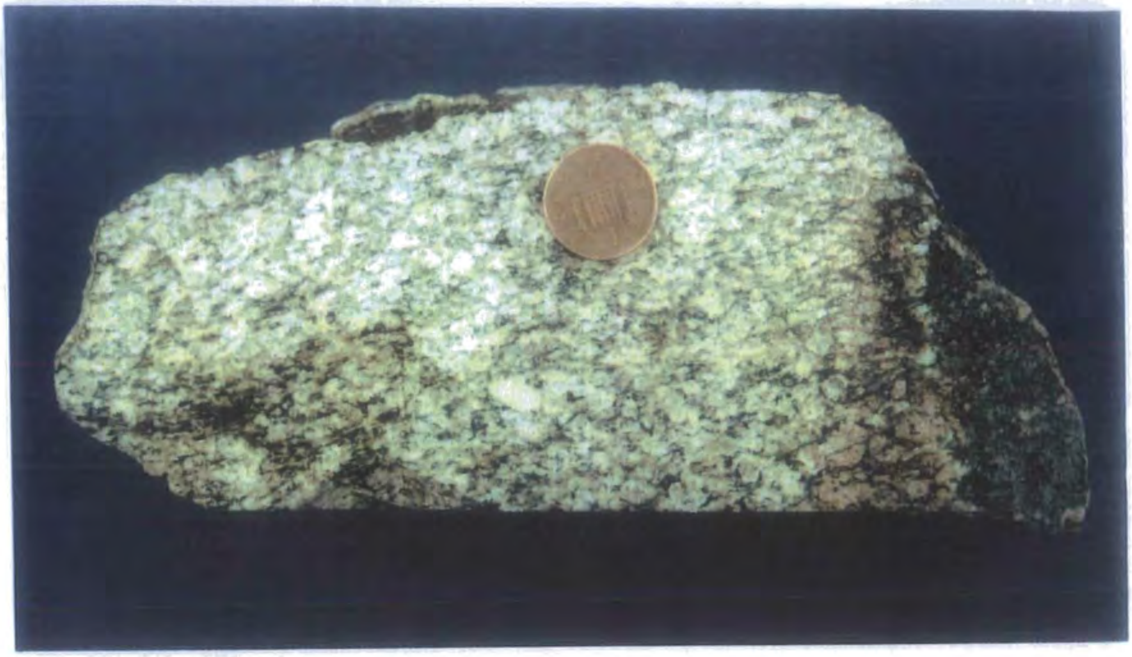
5) **GM 5**:- Coarse-grained pink porphyritic monzo/syenogranite.



6) **GM 6**:- coarse-grained white porphyritic monzogranite.

D) The granitoids of the Barnes Gap

i) The Crockmore apophysis



1) Medium-grained, biotite-rich equigranular monzogranite.

2) Less deformed Crockmore Granite (taken from the rail-section through the Barnes Gap (GR C 093242).

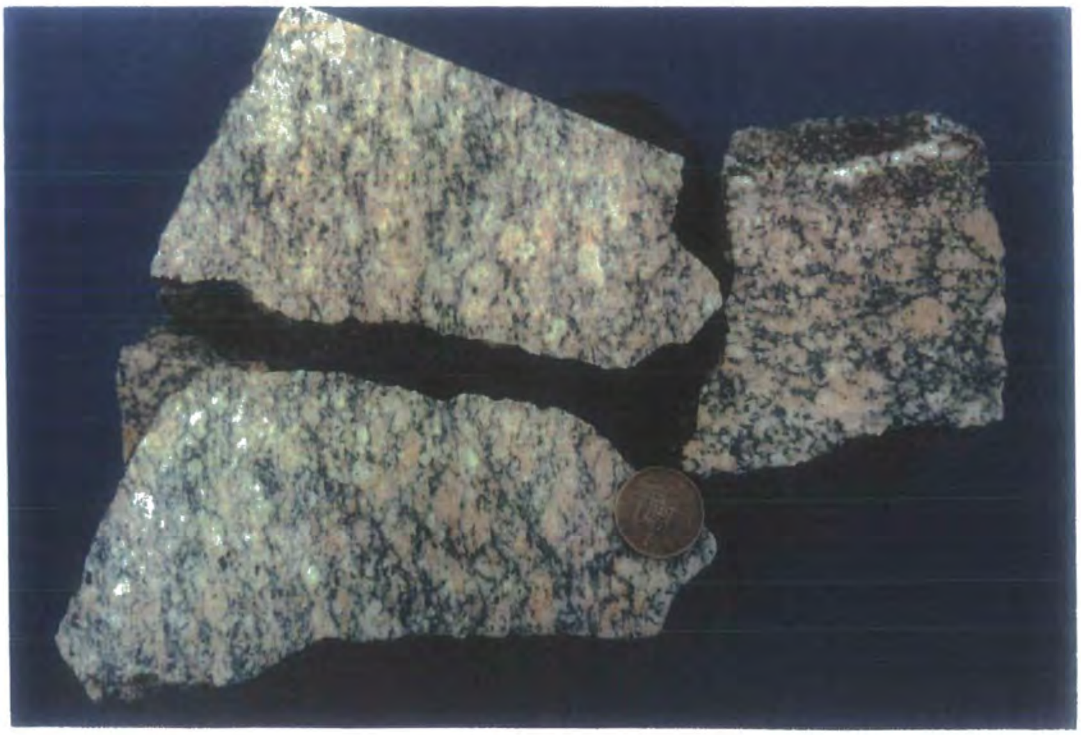
ii) The Binaniller apophysis

1) Pink, subtly porphyritic monzogranite:- this coarser variety is commonly seen within the more central areas of the Binaniller apophysis (B.a. 2) (GR C 095256).

2) White, biotite-rich equigranular granodiorite-monzogranite:- seen at the top and the base of the Binaniller apophysis (B.a. 1) (GR C 091246).

E) The Pink Porphyritic Monzogranites of the NW region of the pluton

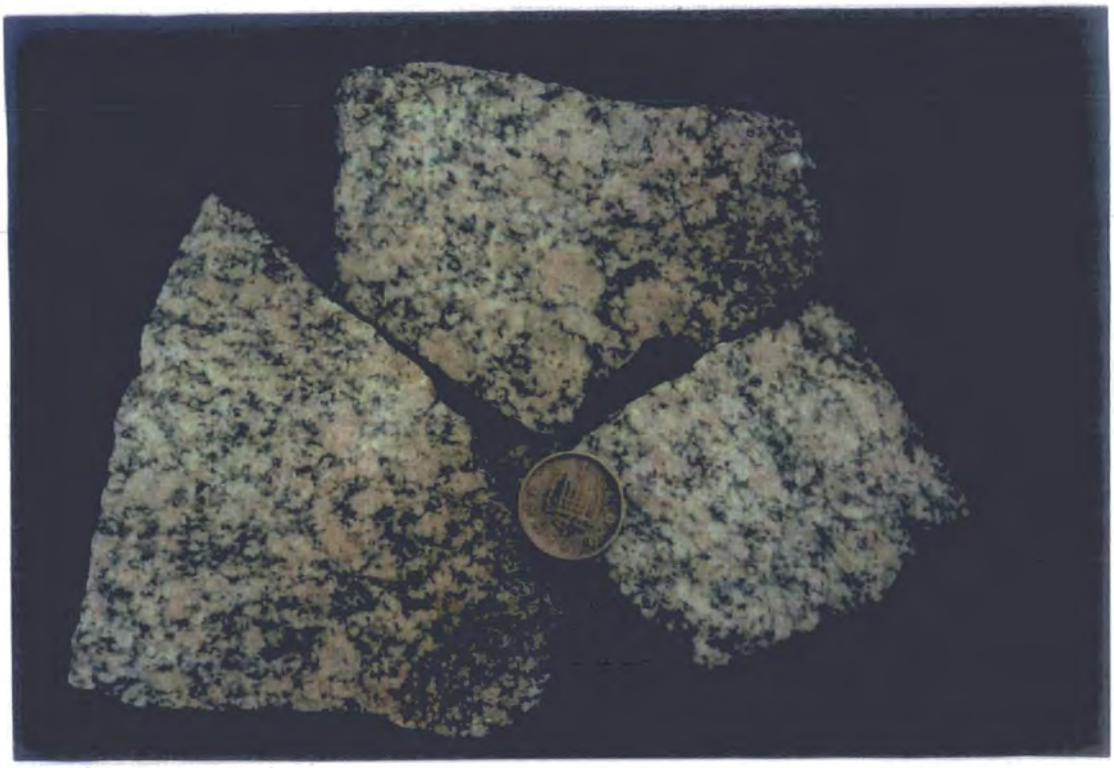
1) Medium to coarse-grained pink porphyritic monzogranite (Glen Granite Quarry GR C 112307).



2) Pink porphyritic monzogranites from other areas along the NW Margin:- *Top left*; Poisoned Glen (GR B 934172), *Top right*; Meenderryherk (GR B 834095), *Bottom right*; Lackagh Bridge (GR C 112307).

F) The Granitoids of the Galwollie Hill area (Map C)

1) **GWH 1**:- Fine to medium-grained equigranular tonalite.

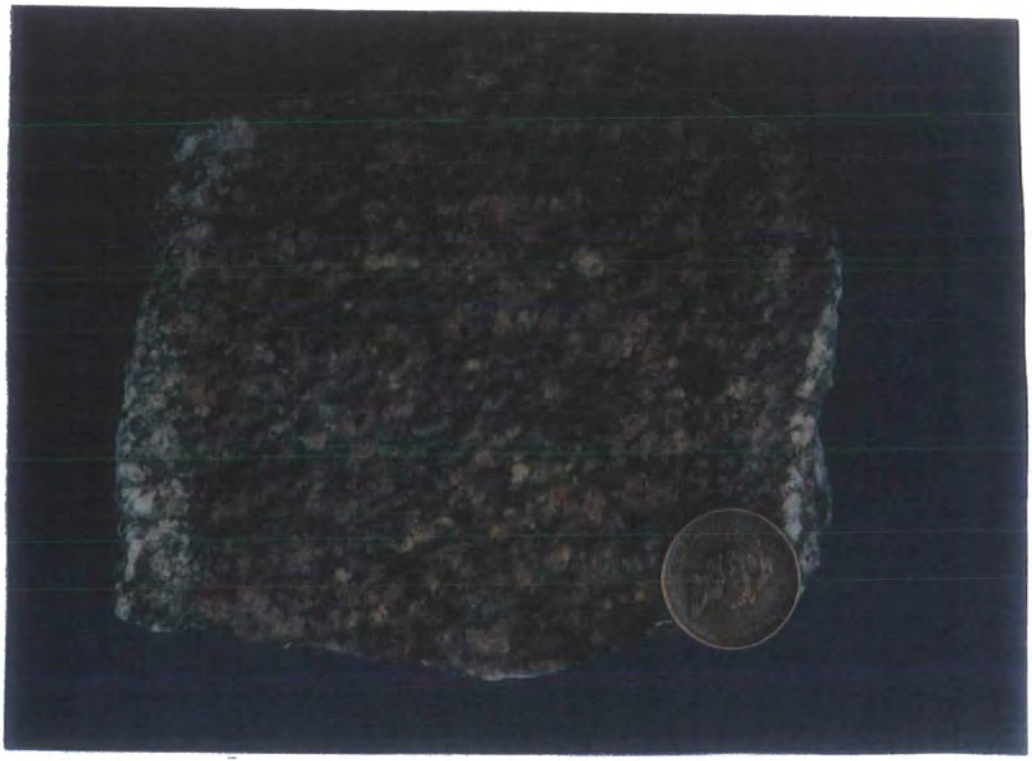


2) **GWH 2**:- medium-grained porphyritic monzogranite.

3) **GWH 3**:- pink-white subtly porphyritic monzogranite.

4) **G II (Trawenagh Bay Granite)**. North of Trusklieve Hill (GR B 907014).

5) **GWH 4**:- medium-grained, subtly porphyritic monzogranite (muscovite-rich). G II of the Trawenagh Bay Granite.



6) **GWH 5**:- red, highly deformed, subtly porphyritic monzogranite (muscovite-rich). (G II of Trawenagh Bay Granite).

Appendix B

Detailed field slips of the areas mapped at a scale of 1:250 giving the locations of where thin-sections and geochemical samples were obtained from

In appendix C & D there are data which refer to thin-section modal analysis and also geochemical analysis. The majority of these samples were obtained form the areas of detailed mapping within this thesis; Map A, Map D, Map C and Maps I & II. This appendix displays the original field slips from these areas and the locations from where all the samples (i.e. thin-sections, geochemical samples and the hand specimens displayed in Appendix A) were obtained.

Appendix C

Modal analyses for all granitoids which occur within the Main Donegal Granite

In this appendix the modal proportions of mineral phases have been obtained by point counting of rock slabs. In some of the K-feldspar-rich granitoids staining of the feldspars was used using sodium cobaltinitrate. The renormalisation values of plagioclase, K-feldspar and quartz to 100 % which allows QAP diagrams to be plotted are also listed in this appendix.

Appendix C: Modal analyses for all granitoids which occur within the Main Donegal Granite

SAMPLE	GRANITE TYPE	LOCATION	N	PLAGIO-CLASE	K-FELDSPAR	QUARTZ	BIOTITE	HORN-BLENDE	MUSCO-VITE	EPIDOTE	OTHERS	MAFIC %
CBH4	CBH 3	MAP A	N=400	139	102	128	18	0	11	0	2	
			%	34.7	25.5	32	4.5	0	2.7	0	0.5	
			QAP	37.50%	27.50%	35.00%						7.7
CBH5	CBH 1	MAP A	N=400	216	11	106	61	0	4	2	0	
			%	54	2.7	26.5	15.2	0	1	0.5	0	
			QAP	64.90%	3.30%	31.80%						16.7
CBH11	CBH 4	MAP A	N=400	204	0	119	63	0	8	5	1	
			%	51	0	29.8	15.7	0	2	1.2	0.2	
			QAP	63.20%	0%	36.80%						19.1
CBH16	CBH 1	MAP A	N=4OO	189	64	96	39	0	8	4	0	
			%	47.2	16	24	9.7	0	2	1	0	
			QAP	54.10%	18.30%	27.60%						12.7
CBH9	CBH 5	MAP A	N=400	126	96	115	40	0	21	0	2	
			%	31.5	24	28.7	10	0	5.2	0	0.5	
			QAP	37.40%	28.50%	34.10%						15.7
CBH14	CBH 2	MAP A	N=400	130	115	100	43	0	12	0	0	
			%	32.5	28.8	25	10.7	0	3	0	0	
			QAP	37.70%	33.30%	29.00%						13.7
CBH8	CBH 2	MAP A	N=400	169	65	132	32	0	1	0	1	
			%	42.25	16.25	33	8	0	0.25	0	0.25	
			QAP	46.20%	17.80%	36.00%						8.5
CBH1	CBH 1	MAP A	N=600	272	94	162	56	0	15	1	0	
			%	45.3	15.6	27	9.3	0	2.5	0.1	0	
			QAP	51.50%	17.70%	30.80%						11.9
CBH3	CBH 4*	MAP A	N=600	310	2	198	80	0	4	6	0	
			%	51.6	0.3	33	13.3	0	0.6	1	0	
			QAP	60.70%	0.40%	38.90%						14.9
CBH21	CBH 4	MAP A	N=600	307	9	176	100	0	2	4	2	
			%	51.1	1.5	29.3	16.6	0	0.3	0.6	0.3	
			QAP	62.30%	1.90%	35.80%						17.8
CBH19	CBH 1	MAP A	N=600	188	127	220	63	0	2	0	0	
			N=600	192	123	209	71	0	5	0	0	
			%	31.3	21.1	36.6	10.5	0	0.3	0	0	
			%	32	20.5	34.8	11.8	0	0.8	0	0	
			QAP	35.20%	23.70%	41.10%						10.8
			QAP	36.60%	23.50%	39.90%						12.6
CBH2	CBH 2	MAP A	N=600	258	119	174	42	0	7	0	0	
			%	43	19.8	29	7	0	1.1	0	0	
			QAP	46.80%	21.60%	31.60%						8.1
CBH22	CBH 2	MAP A	N=600	200	192	182	23	0	3	0	0	
			%	33.3	32	30.3	3.8	0	0.5	0	0	
			QAP	34.80%	33.50%	31.70%						4.3
CBH7	CBH 3	MAP A	N=600	221	174	181	19	0	5	0	0	
			%	36.8	29	30.2	3.1	0	0.8	0	0	
			QAP	38.30%	30.20%	31.50%						3.9
CBH18	CBH 2	MAP A	N=600	184	186	175	48	0	7	0	0	
			%	30.6	31	39.1	8	0	1.1	0	0	
			QAP	33.80%	34.20%	32.00%						9.1
CBH10	CBH 5	MAP A	N=600	181	97	256	56	0	7	3	0	
			N=600	194	102	251	45	0	6	2	0	
			%	30.1	16.1	42.7	9.3	0	1.1	0.5	0	
			%	32.3	17	41.8	7.5	0	1	0.3	0	
			QAP	33.80%	18.20%	48.00%						10.9
			QAP	35.40%	18.60%	46.00%						8.8
CBH20	TONALITE	MAP A	N=700	313	48	194	133	0	11	0	1	
			%	44.7	6.8	27.7	19	0	1.5	0	0.1	
			QAP	56.50%	8.60%	34.90%						20.6
H6H	ARDARA (GI)	B 795954	600	290	16	172	101	13	0	3	5	
			%	48.3	2.7	28.6	16.8	2.2	0	0.6	0.8	
			QAP	60.70%	3.40%	35.90%						20.4
M6V	ARDARA (GI)	B 798934	600	276	30	160	109	12	0	4	9	
			%	46	5	26.6	18.2	2	0	0.7	1.5	
			QAP	59.30%	6.40%	34.30%						22.4
CG3	ARDARA	B 911035	600	255	42	172	129	0	0	0	2	
	CARBAT		%	42.5	7	28.6	21.5	0	0	0	0.3	
			QAP	54.40%	9.00%	36.60%						21.8
CG9	ARDARA	B 908032	700	289	51	210	126	20	0	0	4	
	CARBAT		%	41.3	7.2	30	18	2.8	0	0	0.6	
			QAP	52.60%	9.20%	38.20%						21.5
CARBANE	TONALITE	B 824933	700	337	15	179	134	0	0	33	2	
			%	48.1	2.2	25.6	19.1	0	0	4.7	0.3	
			QAP	63.40%	2.90%	33.70%						24.1
CG2	TONALITE	B 911033	700	284	17	243	99	24	0	28	5	
	(CARBAT)		%	40.6	2.4	34.7	14.2	3.4	0	4	0.7	
			QAP	52.30%	3.10%	44.60%						22.3
K10	THORR	C 040160	700	269	87	171	171	0	0	2	0	
			%	38.4	12.4	24.4	24.4	0	0	0.4	0	
			QAP	51.10%	16.50%	32.40%						24.8

SAMPLE	GRANITE TYPE	LOCATION	N	PLAGIO-CLASE	K-FELDSPAR	QUARTZ	BIOTITE	HORN-BLENDE	MUSCO-VITE	EPIDOTE	OTHERS	MAFIC %
CH4	TONALITE	C 125328	700	414	12	124	143	0	2	1	4	
			%	59.1	1.7	17.7	20.4	0	0.5	0.1	0.5	
			QAP	75.30%	2.20%	22.50%						21.5
CH1	TONALITE	C 125327	700	341	33	190	126	0	5	2	3	
	(THORR)		%	48.7	4.7	27.1	18	0	0.7	0.2	0.4	
			QAP	60.50%	5.80%	33.70%						19.3
LB8	TONALITE	C 096310	700	225	154	178	136	0	0	2	5	
	(THORR)		%	32.2	22.1	25.4	19.4	0	0	0.3	0.8	
			QAP	40.50%	27.80%	31.70%						20.5
CBH20	TONALITE	MAP B	N=700	313	48	194	133	0	11	0	1	
(THORR)	(THORR)		%	44.7	6.8	27.7	19	0	1.5	0	0.1	
			QAP	56.50%	8.60%	34.90%						20.6
CS19	B.a. 2	C 090252	700	216	193	238	48	0	0	5	0	
			%	30.8	27.6	34	6.8	0	0	0.8	0	
			QAP	33.30%	29.90%	36.80%						7.6
CS15	B.a. 2	C 096258	700	240	193	214	41	0	11	1	0	
			%	34.3	27.6	30.5	5.8	0	1.6	0.2	0	
			QAP	37.10%	29.90%	33.00%						7.6
CS22	CG	C 093242	700	206	214	227	49	0	4	0	0	
			%	29.4	30.6	32.4	7	0	0.6	0	0	
			QAP	31.80%	33.10%	35.10%						7.6
BGAP	B.a.3	C 086257	700	239	144	265	51	0	0	1	0	
			%	34.1	20.6	37.8	7.3	0	0	0.2	0	
			QAP	36.90%	22.30%	40.80%						7.5
CS1	CG	C 102259	600	178	144	210	52	0	16	0	0	
			%	29.6	24.2	35	8.6	0	2.6		0	
			QAP	33.30%	27.20%	39.50%						11.2
CS5	CG	C 094249	600	198	125	210	43	0	24	0	0	
			%	33	20.9	35	7.1	0	4	0	0	
			QAP	37.10%	23.50%	39.40%						11.1
K5	CG	C 024165	600	205	98	234	56	0	7	0	0	
			%	34.2	16.3	39	9.3	0	1.2	0	0	
			QAP	38.20%	18.20%	43.60%						10.5
CS18	B.a. 2	C 089252	700	209	219	206	64	0	0	2	0	
			%	29.8	31.3	29.4	9.2	0	0	0.3	0	
			QAP	32.90%	32.50%	32.50%						9.5
CH6	PINK PORP	C 113308	600	166	211	188	31	0	4	0	0	
			%	27.6	35.2	31.3	5.3	0	1.6	0	0	
			QAP	29.30%	37.40%	33.30%						5.9
PG4	PINK PORP	B936167	700	191	267	196	30	0	15	1	0	
			%	27.3	38.1	28	4.2	0	2.2	0.2	0	
			QAP	29.30%	40.80%	29.90%						6.6
TB10	PINK PORP	B 837092	700	303	199	159	35	0	2	2	0	
			%	43.2	28.4	22.7	5	0	0.4	0.3	0	
			QAP	45.80%	30.10%	24.10%						5.7
DS2	MONZO	B 864074	700	238	182	219	47	0	13	1	0	
			%	34	26	31.3	6.7	0	1.8	0.2	0	
			QAP	37.20%	28.50%	34.30%						8.7
DS4	MONZO	B 864074	700	211	215	256	10	0	8	0	0	
			%	30.1	30.7	36.6	1.4	0	1.2	0	0	
			QAP	30.90%	31.50%	37.60%						2.6
DS8	GRANODIO	B 867072	600	281	43	202	61	0	9	4	0	
			%	46.8	7.1	33.6	10.2	0	1.6	0.7	0	
			QAP	53.50%	8.10%	38.40%						12.5
CH3	TONALITE	C 125327	700	356	1	213	114	0	5	10	1	
			%	50.8	0.2	30.4	13.6	0	0.7	1.6	0.1	
			QAP	62.50%	0.00%	37.50%						18.6
LB7	FINE DYKE	C 096310	700	220	54	298	109	0	3	15	1	
			%	31.4	7.7	42.5	15.5	0	0.4	2.1	0.4	
			QAP	38.50%	9.40%	52.10%						18.4
CG4	APLOMDG	B 910034	700	237	159	271	28	0	5	0	0	
			%	33.8	22.7	38.8	4	0	0.7	0	0	
			QAP	35.50%	23.80%	40.70%						4.7
CG7	"GM 1"	B 935067	600	331	24	159	86	0	0	0	0	
			%	55.2	4	26.5	14.3	0	0	0	0	
			QAP	64.40%	4.70%	30.90%						14.3
CU 10	GM 1	MAP I	N=600	274	54	191	80	0	0	1	0	
			%	45.7	9	31.8	13.3	0	0	0.2	0	
			QAP	52.80%	10.40%	36.80%						13.5
CU 1	GM 2	MAP I	N=600	288	14	228	69	0	0	1	0	
			%	48	2.3	38	11.5	0	0	0.2	0	
			QAP	54.40%	2.60%	43.00%						11.7
CU 21	GM 2	MAP II	N=600	334	0	184	71	0	10	1	0	
			%	55.6	0	30.8	11.8	0	1.6	0.2	0	
			QAP	64.40%	0%	35.60%						13.6
CU 22	GM 2	MAP II	N=600	323	19	183	61	0	14	0	0	
			%	53.8	3.1	30.6	10.2	0	2.3	0	0	
			QAP	61.50%	3.50%	34.00%						12.5

Appendix C: Modal analyses for all granitoids which occur within the Main Donegal Granite

SAMPLE	GRANITE TYPE	LOCATION	N	PLAGIO-CLASE	K-FELDSPAR	QUARTZ	BIOTITE	HORN-BLENDE	MUSCO-VITE	EPIDOTE	OTHERS	MAFIC %
CU 18	GM 6	C 965135	N=600	252	114	183	45	0	6	0	0	
			N=600	221	102	216	57	0	4	0	0	
			N=600	210	115	218	54	0	3	0	0	
			%	42	19.2	30.6	7.2	0	1	0	0	
			%	36.8	17.1	36	9.5	0	0.6	0	0	
			%	35	19.2	36.3	9	0	0.5	0	0	
			QAP	45.80%	20.90%	33.50%						8.2
			QAP	40.90%	19.00%	40.10%						9.5
			QAP	38.70%	21.20%	40.10%						10.1
CU 19	GM 6	MAP II	N=600	180	217	167	34	0	2	0	0	
			N=600	210	133	215	38	0	4	0	0	
			%	30	36.2	27.8	5.7	0	0.3	0	0	
			%	35	22.2	35.8	6.4	0	0.6	0	0	
			QAP	31.90%	38.50%	29.60%						6
			QAP	37.60%	23.90%	38.50%						7
		AVER	QAP	34.70%	31.80%	33.50%						6.2
M 3	GM 5	C 951135	N=600	239	149	166	21	0	25	0	0	
			%	39.8	24.8	27.6	3.6	0	4.2	0	0	
			QAP	43.20%	26.90%	29.90%						7.8
CU 7	GM 4	MAP I	N=700	199	249	217	30	0	4	0	1	
			%	28.5	35.6	31	4.2	0	0.5	0	0.2	
			QAP	30.00%	37.40%	32.60%						4.9
M 2	GM 4	B 951134	N=700	252	194	198	44	0	11	0	1	
			%	36	27.7	28.3	6.3	0	1.5	0	0.2	
			QAP	39.10%	31.10%	30.80%						8
CU 6	GM 3	MAP I	N=700	247	196	207	43	0	4	3	0	
			%	35.3	28	29.5	6.1	0	0.6	0.5	0	
			QAP	38.00%	30.20%	31.80%						7.2
CU 3	GM 6	MAP I	N=700	143	349	181	18	0	9	0	0	
			%	20.6	49.8	25.8	2.6	0	1.3	0	0	
			QAP	21.40%	51.80%	26.80%						3.9
SA188	SRU 1	MAP B	N=500	213	101	95	80	0	10	1	0	
			%	42.6	20.2	19	16	0	2	0.2	0	
			QAP	52.10%	24.70%	23.20%						18.2
SA6	SRU 1	MAP B	N=500	242	81	98	69	0	8	2	0	
			%	48.4	16.2	19.6	13.8	0	1.6	0.4	0	
			QAP	57.50%	19.20%	23.30%						15.8
SA39	SRU 1	MAP B	N=500	207	122	95	67	0	3	6	0	
			%	41.4	24.4	19	13.4	0	0.6	1.2	0	
			QAP	48.80%	28.80%	22.40%						15.2
SA242	SRU 1	SUB-MAP 10	N=600	209	127	164	100	0	0	0	0	
			%	34.4	21.1	27.3	16.6	0	0	0	0	
			QAP	41.70%	25.60%	32.70%						16.6
SA19	SRU 2	MAP B	N=500	300	12	116	61	0	9	2	0	
			%	60	2.4	23.2	12.2	0	1.8	0.4	0	
			QAP	71.00%	2.80%	27.10%						14.4
SA86	SRU 2	MAP B	N=500	252	1	162	74	0	8	3	0	
			%	50.4	0.2	32.4	14.8	0	1.6	0.6	0	
			QAP	60.70%	0.30%	39.00%						17
SA65	SRU 2	MAP B	N=500	251	1	166	76	0	3	3	0	
			%	50.2	0.2	33.2	15.2	0	0.6	0.6	0	
			QAP	60.00%	0.30%	39.70%						16.4
SA177	SRU 2	MAP B	N=500	267	0	146	78	0	9	0	0	
			%	53.4	0	29.2	15.6	0	1.8	0	0	
			QAP	64.70%	0%	35.50%						17.4
SA3	SRU 2	MAP B	N=500	278	4	159	50	0	9	0	0	
			%	55.6	0.8	31.8	10	0	1.8	0	0	
			QAP	63.00%	0.90%	36.10%						11.8
SA8	SRU 2	MAP B	N=600	269	0	225	97	0	9	0	0	
			%	44.8	0	37.5	16.1	0	1.8	0	0	
			QAP	54.50%	0%	45.50%						17.9
SA179	SRU 2	MAP B	N=600	280	28	180	103	0	8	1	0	
			%	46.7	4.7	30	17.1	0	1.3	0.1	0	
			QAP	57.20%	6.00%	36.80%						18.5
SA62	SRU 2	MAP B	N=500	241	0	183	67	0	8	1	0	
			%	48.2	0	36.6	13.2	0	1.6	0.2	0	
			QAP	56.80%	0%	43.20%						15.2
SA29	DYKE	MAP B	N=500	135	153	173	34	0	3	2	0	
			%	27	30.6	34.6	6.8	0	0.6	0.4	0	
			QAP	29.30%	33.20%	37.50%						7.8
SA4	DYKE	MAP B	N=500	133	179	140	41	0	7	0	0	
			%	26.6	35.8	28	8.2	0	1.4	0	0	
			QAP	29.40%	39.60%	31.00%						9.6
SA202	DYKE	MAP B	N=500	204	65	161	62	0	8	0	0	
			%	40.8	13	32.2	12.4	0	1.6	0	0	
			QAP	47.50%	15.10%	37.40%						14
SA136	SRU 3: PSBG	MAP B	N=600	206	173	148	64	0	4	5	0	
			%	34.3	28.8	24.6	10.6	0	0.6	0.8	0	
			QAP	39.20%	32.70%	28.10%						12

SAMPLE	GRANITE TYPE	LOCATION	N	PLAGIO-CLASE	K-FELDSPAR	QUARTZ	BIOTITE	HORN-BLENDE	MUSCO-VITE	EPIDOTE	OTHERS	MAFIC %
SA5	SRU 3: PSBG	MAP B	N=600	191	207	138	51	0	12	1	0	
			%	31.8	34.5	23	8.5	0	2	0.1	0	
			QAP	35.60%	38.60%	25.80%						10.6
SA184	SRU 3: PPBG	MAP B	N=600	198	196	154	46	0	5	1	0	
			%	33	32.7	25.7	7.7	0	0.8	0.1	0	
			QAP	36.30%	35.30%	28.40%						8.6
SA96	SRU 3: PSBG	MAP B	N=600	212	156	174	51	0	6	1	0	
			%	35.3	26	29	8.5	0	1	0.1	0	
			QAP	39.10%	28.80%	32.10%						9.6
SA107	SRU 3: PSBGp	MAP B	N=600	227	138	159	57	0	16	3	0	
			%	37.8	23	26.5	9.5	0	2.6	0.5	0	
			QAP	43.30%	26.40%	30.30%						12.6
CWBG	SRU 3:CWBG	SUB-MAP 11	N=600	227	170	170	28	0	3	2	0	
			%	37.8	28.3	28.3	4.6	0	0.5	0.3	0	
			QAP	40.20%	29.90%	29.90%						5.4
SA16	SRU 3:VCPG	MAP B	N=650	196	211	200	30	0	13	0	0	
			%	30.2	32.5	30.8	4.6	0	2	0	0	
			QAP	32.30%	34.70%	33.00%						6.6
SA247	SRU 3: PPBG	SUB-MAP 10	N=700	223	266	173	31	0	4	3	0	
			%	31.8	38	24.7	4.4	0	0.4	0.4	0	
			QAP	33.60%	40.20%	26.20%						5.3
SA95	SRU 3: PPBG	MAP B	N=700	266	252	141	27	0	13	1	0	
			%	38	36	20.1	3.8	0	1.8	0.1	0	
			QAP	40.30%	38.30%	21.40%						5.7
SA196	SRU 3: PSBG	MAP B	N=600	169	221	144	50	0	15	1	0	
			%	28.2	36.8	24	8.3	0	2.4	0.1	0	
			QAP	31.70%	41.40%	26.90%						10.8
SA156	SRU 3: PSBG p	MAP B	N=600	204	187	165	35	0	6	1	2	
			%	34	31.2	27.5	5.6	0	1	0.1	0.3	
			QAP	36.70%	33.60%	29.70%						7
SA237	SRU 3: VCPG	SUB-MAP 9	N=700	234	213	202	39	0	12	0	0	
			%	33.4	30.4	28.8	5.5	0	1.7	0	0	
			QAP	36.00%	32.80%	31.20%						7.2
SA219	DYKE	SUB-MAP 9	N=400	139	0	127	134	0	0	0	0	
			N=500	243	0	111	146	0	0	0	0	
			%	34.7	0	31.7	33.5	0	0	2	0	
				48.6	0	22.2	29.2	0	0	0.4	0	
			QAP	52.20%	0%	47.80%						33.5
			QAP	67.80%	0%	32.20%						29.2
9GH3	GWH 1	MAP C	600	316	14	170	89	0	8	2	1	
			%	52.6	2.3	28.3	14.9	0	1.3	0.4	0.2	
			QAP	63.20%	2.80%	34.00%						16.8
2GH3	GWH 1	MAP C	600	336	3	187	68	0	5	1	0	
			%	56	0.5	31.2	11.3	0	0.8	0.2	0	
			QAP	63.90%	0.60%	35.50%						12.3
4GH3	GWH 1	MAP C	700	380	4	232	74	0	9	0	1	
			%	54.2	0.6	33.2	10.5	0	1.3	0	0.2	
			QAP	61.60%	0.70%	37.70%						12
BGH3	GWH 1	MAP C	700	359	16	237	87	0	0	1	0	
			%	51.2	2.3	33.8	12.5	0	0	0.2	0	
			QAP	58.70%	2.60%	38.70%						12.7
2CGH4	GWH 2	MAP C	700	271	153	206	60	0	10	0	0	
			700	261	207	179	48	0	5	0	0	
			%	38.7	21.9	29.4	8.5	0	1.4	0	0	
			%	37.3	29.5	25.6	6.8	0	0.7	0	0	
			QAP	43.00%	24.30%	32.70%						10
			QAP	40.30%	32.00%	27.70%						7.5
E1	GWH 3: GII	MAP C	700	246	140	267	37	0	10	0	0	
			%	35.1	20	38.2	5.3	0	1.4	0	0	
			QAP	37.60%	21.40%	41.00%						6.7
C>GH3	GWH 4	MAP C	550	138	155	234	18	0	3	0	2	
			%	25.1	28.2	42.6	3.2	0	0.5	0	0.4	
			QAP	26.20%	29.40%	44.40%						4.1
S1	GWH 4	MAP C	700	183	249	231	29	0	9	0	1	
			%	26.1	35.5	33	3.8	0	1.3	0	0.2	
			QAP	27.60%	37.60%	34.80%						5.3
5CGH30	GWH 5	MAP C	700	170	203	290	26	0	11	0	0	
			%	24.3	29	41.4	3.7	0	1.6	0	0	
			QAP	25.70%	30.60%	43.70%						5.3
1CGH10	GWH 5	MAP C	700	173	185	278	43	0	19	0	2	
			%	24.7	26.4	39.8	6.2	0	2.8	0	0.3	
			QAP	27.20%	29.10%	43.70%						9.2
6CGH20	GWH 5	MAP C	700	191	192	251	36	0	30	0	0	
			%	27.3	27.4	35.8	5.2	0	4.3	0	0	
			QAP	30.20%	30.30%	39.50%						9.5
CGH3	GWH 4	MAP C	700	186	224	227	34	0	29	0	0	
			%	26.6	32	32.4	4.8	0	4.2	0	0	
			QAP	29.20%	35.20%	35.60%						9

SAMPLE	GRANITE TYPE	LOCATION	N	PLAGIO-CLASE	K-FELDPSAR	QUARTZ	BIOTITE	HORN-BLENDE	MUSCO-VITE	EPIDOTE	OTHERS	MAFIC %
3CGH25	GWH 5	MAP C	700	198	251	192	29	0	30	0	0	
			%	28.3	35.8	27.5	4.1	0	4.3	0	0	
			QAP	30.90%	39.10%	30.00%						8.4
4CGH10	GWH 5	MAP C	700	178	214	270	9	0	28	0	1	
			%	25.4	30.5	38.6	1.3	0	4	0	0.2	
			QAP	26.90%	32.30%	40.80%						5.5
2CGH1	GWH 5	MAP C	700	150	246	264	20	0	19	1	0	
			%	21.4	35.2	37.7	2.8	0	2.7	0.2	0	
			QAP	22.70%	37.30%	40.00%						5.7
SEG	GWH 3	MAP C	700	240	178	219	46	0	15	0	2	
			%	34.3	25.4	31.3	6.5	0	2.2	0	0.3	
			QAP	37.70%	27.90%	34.40%						9
3CGH12	GWH 3	MAP C	700	248	189	209	40	0	13	1	0	
			%	35.4	27	29.8	5.7	0	1.9	0.2	0	
			QAP	38.40%	29.30%	32.30%						7.8
L2	GWH 5	MAP C	700	159	202	298	23	0	18	0	0	
			%	22.7	28.8	42.6	3.3	0	2.6	0	0	
			QAP	24.10%	30.60%	45.30%						5.9
2CGH2	GWH 5	MAP C	700	220	195	238	27	0	18	2	0	
			%	31.4	27.8	34	3.9	0	2.6	0.3	0	
			QAP	33.70%	29.80%	36.50%						6.8
CGH9	GWH 3	MAP C	700	208	173	258	48	0	12	1	0	
			%	29.7	24.7	36.8	6.8	0	1.8	0.2	0	
			QAP	32.60%	27.10%	40.30%						8.8
3CGH13	GWH 5	MAP C	700	192	191	283	23	0	11	0	0	
			%	27.4	27.3	40.4	3.3	0	1.6	0	0	
			QAP	28.80%	28.70%	42.50%						4.9
L1	GWH 1	MAP C	700	280	12	294	82	0	28	4	0	
			%	40	1.7	42	11.7	0	4	0.6	0	
			QAP	47.80%	2.00%	50.20%						16.3
TB3	G I	B 907014	700	238	171	209	71	0	7	3	1	
			%	34	24.4	29.8	10.2	0	1	0.4	0.2	
			QAP	38.60%	27.70%	33.70%						11.8
1CGH6	GWH 5	MAP C	700	191	220	262	6	0	21	0	0	
			%	27.3	31.4	37.4	0.9	0	3	0	0	
			QAP	28.40%	32.70%	38.90%						3.9
S3 (b)	GWH 5	MAP C	700	210	199	230	47	0	14	0	0	
			%	30	28.4	32.9	6.7	0	2	0	0	
			QAP	32.90%	31.10%	36.00%						8.7
A6	GWH 5	MAP C	600	152	197	208	22	0	17	0	4	
			%	25.3	32.8	34.6	3.6	0	2.9	0	0.8	
			QAP	27.30%	35.40%	37.30%						7.3
TB14	G I	B 834095	700	237	179	232	34	0	9	7	2	
			%	33.8	25.6	33.1	4.9	0	1.3	1	0.3	
			QAP	36.50%	27.70%	35.80%						7.5
TRU	G I	B 790018	700	298	179	168	52	0	3	0	0	
			%	42.5	25.6	24	7.4	0	0.4	0	0	
			QAP	46.20%	27.80%	26.00%						7.8
TB2	"GWH 2"	B 819023	700	31.8	46	164	69	0	1	1	1	
			%	53	7.6	27.3	11.5	0	0.2	0.2	0.2	
			QAP	60.30%	8.70%	31.00%						12.1

Appendix D

Geochemical data for the granitoids of the Main Donegal Granite

Preparation of rock powders

Hand specimens collected in the field were on average 2-3 kg in size with the majority of the weathered material removed in the field. In the Sruhanavarnis, due to horizontal glacial surfaces a rock drill was used. Some of the geochemical samples were taken from 3-4 rock cores from the same location. On average each of these cores were 10 cm in length. All weathered material was sawn off before or during the splitting of the samples into smaller cuboidal pieces, 3-4mm in diameter. The samples were split using a stainless steel hydraulic splitter. Before crushing the samples were washed with water and cleaned with a bristle brush. The clean, dry rock, pieces were crushed to fine gravel using a "Fritsch Pulverissette" jaw crusher (type 01-704). Between each sample the jaw crusher was cleaned with absolute alcohol and wire brush. To keep dust to a minimum a vacuum cleaner underneath the jaw crusher was used.

The fine gravel was powdered using a tungsten carbide swing mill until a fine powder was formed which usually took 60-90 seconds. After each sample was powdered the swing-mill was first cleaned with water (+ bristle brush) and then with absolute alcohol and thoroughly dried before further use. The use of a tungsten carbide swing mill results in considerable contamination by tungsten, significant cobalt, tantalum and scandium with trace amounts of niobium (Rollinson 1993).

XRF Analysis

Powdered Pellets

Of the powdered samples prepared, approximately 5-6g were added to 7-8 drops of 4% solution of Moviol (organic adhesive) and were subsequently mixed thoroughly. The mixture was transferred into a mould where it was compressed at 10 tons\psi for two minutes. The briquettes were then dried in a oven for 12 hours at 110°C and were also dried the night before analysis. Samples were kept in airtight bags when not being analysed. For coarse grained samples a number of replica samples were made with the average whole rock chemistry for these specimens being calculated.

Fusion Discs

Despite the majority of the major element chemistry having derived from analysis of powdered briquettes some fusion disc analysis was performed. The loss on ignition (LOI) was first calculated by heating 5g of the powder in a furnace at 900°C for two hours. The powder was allowed to cool before reweighing. On average the granite powders lost weight. The LOI ranges for the most mafic granite to the least mafic granite was between 0.442-0.847% H_2O. Of the powder used in LOI 0.25g of it was mixed thoroughly, in an agate pestle and mortar, with Spectroflux 100B (lithium tetraborate flux). This mixture was then placed in platinum crucibles and heated to 1050°C for 20 minutes. The molten glass was then poured into moulds on a hot plate and were immediately flattened into the required shape with a stainless steel plunger. The prevent quench shattering heated beakers were placed over the moulded glass to allow the glass discs to cool more slowly. After cooling the samples were labelled and placed in airtight bags and stored in a dessicator. The platinum crucibles were cleaned in boiling dilute hydrochloric acid for 10 minutes. All essential equipment was cleared with absolute alcohol prior to further sample preparation.

XRF data acquisition

Major element oxides and trace element concentrations (parts per million) were determined using a Philips PW1500 spectrometer using a rhodium anode tube. The initial calibration was performed using a range of compositionally different International standards. The calibration computer programme calculates the best fit straight line of composition versus counts for every element. The operator (Ron Hardy; senior experimental officer) may subsequently reduce the root mean square (RMS) error of the regression line by removing standards that plot away from this line (*taken from* Loughlin 1995).

To test the calibration several International Standards were ran as unknown standards. For major elements these standards were DNC-1, AGV-1 and G-2. As well as international standards several replica pellets were analysed (for major elements) at several stages during the overall analysis as unknown standards to give overall machine precision throughout the analysis. These samples were SA 005, CBH 01, CBH 02, CBH 04 and GH 07. These samples encompass the entire compositional range seen within the granites to be studied. For trace elements the International Standards used were AGV-1, DNC-1, W-2 and G-2, whilst the internal unknown standards were CBH 01, CBH 02, CBH 04, CBH 11, GH 07, SA 005 and SA 023.

A similar method was used for the fusion disc analysis.

Analytical Error

Then errors within sampling were presented in the form of 2σ (2 standard deviations) and are displayed a on the following two pages. Overall the majority of the major oxides determined from powdered pellets were acceptable, although the 2σ values for Al_2O_3 and SiO_2 were more dubious with some samples greater than 1. The trace elements were also generally acceptable although where elements were in quite high concentrations greater 2σ values were encountered, e.g. Ba & Zr. Other elements with quite large 2σ include Ga, Zn, Th and V.

Comment on the Organisation of the geochemical sampling listings

The majority of the geochemical samples were collected in four areas of detailed mapping:- Sruhanavarnis "SA", Crobane Hill "CBH", Glendowan Mountains "GM" and Galwollie Hill "GWH". If the samples from these areas have no grid reference then they can be found in either Map B, Map A, Maps I and II, or Map C of this thesis. In appendix B are photocopies of the detailed field sheets which comprise these areas and these indicate the exact location from where these samples were obtained. For all the other geochemical samples collected from the pluton these samples will be given six-figure grid references to locate them.

	G2					**DNC-1**					**AGV-1**			
	Value	**Average**	**2 s.d.**			**Value**	**Average**	**2 s.d.**			**Value**	**Average**	**2 s.d.**	
		N=4					N=3					N=3		
SiO2	69.04	68.69	0.838			47.04	47.49	0.8096			59.25	59.73	1.386	
Al2O3	15.14	15.14	0.3736			18.3	18.6	0.5922			17.15	16.97	0.416	
K2O	4.49	4.437	0.132			0.229	0.234	0.0176			2.9	2.86	0.078	
Na2O	4.07	4.14	0.1692			1.87	1.84	0.063			4.25	4.1	0.311	
Fe2O3	2.67	2.63	0.11			9.93	10.32	0.8378			6.76	6.91	0.302	
CaO	1.97	1.94	0.092			11.27	11.42	0.306			4.94	4.96	0.045	
MgO	0.76	0.87	0.232			10.05	10.13	0.1612			1.53	1.135	0.79	
TiO2	0.492	0.485	0.0188			0.48	0.48	0			1.06	1.067	0.012	
P2O5	0.14	0.137	0.0149			0.085	0.0845	0.004			0.48	0.458	0.049	
MnO	0.034	0.031	0.006			0.149	0.155	0.013			0.096	0.1035	0.015	
	SA 005			**CBH 01**			**CBH 2**			**CBH 04**			**GH 07**	
	Average	**2 s.d**		**Average**	**2 s.d**		**Average**	**2 s.d**		**Average**	**2 s.d**		**Average**	**2 s.d**
	N=3			N=3			N=3			N=3			N=3	
SiO2	68.89	0.624		69.78	0.586		70.72	1.25		73.39	0.504		67.87	1.438
Al2O3	16.44	0.784		15.21	0.768		15.81	1.17		14.76	1.272		16.68	1.376
K2O	4.667	0.034		2.633	0.034		4.148	0.054		4.63	0.14		2.432	0.026
Na2O	3.85	0.201		4.15	0.184		3.98	0.294		3.58	0.32		4.2	0.214
Fe2O3	2.11	0.08		2.34	0.08		1.96	0.214		0.97	0.034		2.85	0.088
CaO	1.93	0.02		2.77	0.06		2.23	0.061		1.49	0		3.437	0.03
MgO	0.81	0.02		0.986	0.031		0.62	0.064		0.25	0.064		1.45	0.02
TiO2	0.302	0.006		0.323	0.002		0.246	0.024		0.1155	0.001		0.406	0.004
P2O5	0.148	0.008		0.114	0.006		0.12	0.014		0.039	0.014		0.158	0.008
MnO	0.024	0		0.039	0.002		0.293	0.004		0	0		0.487	0.001

	G-2			AGV-1			DNC-1			W-2		
	Value	Average	2 s.d.	Value	Average	2 s.d.	Value	Average	2 s.d.	Value	Average	2 s.d.
		N=3			N=3			N=3				
Nb	13	12	3.28	15	15.2	1.08	3	3.3	0.51	7.9	7.2	1.77
Zr	300	302.3	11.36	225	230.8	11.63	41	36.6	8.81	94	90.4	8.35
Y	11.4	11.7	2.01	21	19	4.05	18	17	2.09	24	21.6	5.58
Sr	478	479	10.32	662	667.9	11.7	145	145.1	0.83	194	195.8	3.65
U	2	1.9	2.19							0.5	1.5	1.73
Th	24.6	26.7	5.41	6.5	4.9	3.26	0.2	1.3	3.27	2.2	2.8	1.87
Pb	31	26.7	8.7	36	34.3	3.42	6.3	8.3	4	9.3	8.5	3
Ga	22	21	2.76	20	22.1	4.37	15	13.4	3.71	20	19	2.06
Zn	85	86.7	4.34	88	86.1	6.93	66	66	1.7	77	78.7	3.5
Cu	11	13.9	6.14	60	58.6	2.86	96	88.3	16.26	103	103.8	7.05
Ni	4.9	5.5	3.29	17	17	0	247	238.1	18.39	70	70	4.06
Co	4.9	6.3	7.26							44	47.9	
Cr	9	6.1	6.37	12	11.1	2.06	285	286.4	2.91	93	91.1	4.14
Rb	170	167.9	4.29	67	67.5	1.56	4.5	5.3	1.61	20	20.5	1.45
Ba	1880	1873	16.57	1221	1222.9	4.47	114	117.9	8.23	182	177.8	11.58
Ce	159	159.9	4.1	66	69.3	72.65	10.6	9.1	3.44	24	25.9	6.28
Nd	53	55.6	6.49	34	35.1	2.58	4.9	5.2	1.21	14	13.4	2.91
La	86	87.3	6.33	38	42.1	8.7	3.8	4.1	2.09	11.4	9	4.92
Sc	3.5	4	1.43	12.1	11.7	1	31	28.7	5.39	35	31.1	8.4
V	36	39.4	7.47	123	120.9	4.24	148	148.6	3.8	262	260.5	6.5

	CBH01		CBH02		CBH04		CBH11		GH07		SA005		SA023	
	Average	2 s.d.	Average	2 s.d.	Average	2 s.d.	Average	2 s.d.	Average	2 s.d.	Average	2 s.d.	Average	2 s.d.
		N=3		N=3		N=3		N=3		N=3		N=4		N=3
Nb	8.5	0.95	9.3	0.23	4.8	1.36	6.3	4.06	9.7	1.74	11.5	1.89	11.9	1.85
Zr	96.1	20.49	137.8	13.28	61.4	15.92	142.6	27.49	128.1	15.21	145.7	20.73	91.8	8.94
Y	8.9	4.35	12.2	5.85	7.4	4.66	7.7	0.61	13.8	2.55	11.8	7.01	17.2	2.8
Sr	368	12.25	356.8	13.93	316.4	8.45	596	60.05	471.8	14.17	447.5	14.45	371.7	9.91
U	1.3		5.1	7.5	4.6	8.12	2.5	0	5.9	3.25	3.8	4.8	5.8	5.15
Th	3.3	2.32	13.1	18.48	12.4	24.76	15.2	30.48	8.3	9.62	9.8	16.64	11.8	4.07
Pb	17.3	3.2	22.2	6.9	22.6	10.9	11.5	3.9	15.3	5.59	24	4.47	16.4	1.85
Ga	19.4	2.25	25.3	24.89	18.2	2.14	23.9	3.78	29.2	33.2	24.1	23.37	21.9	10.5
Zn	49.3	5.61	29.3	38.09	17.8	15.24	59.4	6.08	33.6	42.6	34.8	31.13	22.2	17.49
Cu	6.4	0.99	8.8		8.5		31.4	4.62	9.6		10.4	-1.56		
Ni	7.8	5.62	9	9.6	8	4.02	6.1	14.47	9.5	9.41	7.8	8.34	3.3	5.65
Co	32		48.5		37.7		54.1	11.6	41.1		45.9	0.14	56.7	2.55
Cr	34.7	16	17.3	4.92	12.4	9.11	30.5	17.03	28.1	3.54	20.1	2.74	17.8	1.14
Rb	88.3	0.31	133.3	0.69	115	2.91	84.8	5.44	102.1	3.33	145.7	2.75	69	2.12
Ba	684.3	80.75	1480.2	50.77	1527.6	123.21	420.1	28.66	605.5	58.74	1744.3	83.01	155.2	6.96
Ce	21.9	5.3	37.9	6	23.2	3.97	41.5	5.85	33.1	3.67	43.2	5.78	52.6	1.63
Nd	10.3	1.4	17.8	4.72	7.4	2.69	15	9.36	13.5	9.69	15.9	3.35	20.1	3.37
La	11.7	1.33	24.1	5.9	10.9	5.31	24.9	7.43	21.1	1.22	22.4	1.68	31.1	1.4
Sc	4.4	1.7	4.5	1.79	4	5.13	6.1	0.31	6.9	5.44	4.1	1.04	4.1	1.02
V	34	2.62	17.5	8.55	10.4	9.93	45.8	13.36	44.5	1.7	24.9	4.51	12.1	4.01

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe2O3	TiO2	MnO	CaO	K2O	SiO2	Al2O3	P2O5	MgO	Na2O	TOTAL
Barnes Gap/Crockmore	C 102267	CS09	B.a. 2	2.13	0.304	0.041	2.19	4.256	70.55	15.76	0.154	0.79	4.19	100.36
Barnes Gap/Crockmore	C 099265	CS12	B.a. 3	1.95	0.286	0.039	1.99	4.421	70.77	15.57	0.121	0.79	4.09	100.02
Barnes Gap/Crockmore	C 096261	CS13	B.a. 3	1.82	0.245	0.04	1.68	4.39	72.34	15.56	0.119	0.67	4.35	101.22
Barnes Gap/Crockmore	C 095254	CS14	B.a. 2	1.78	0.237	0.047	1.67	4.342	72.35	15.57	0.11	0.63	4.52	101.25
Barnes Gap/Crockmore	C 094261	CS16	B.a. 2	2.14	0.282	0.045	1.93	4.174	70.85	15.52	0.136	0.75	4.21	100.03
Barnes Gap/Crockmore	C 089282	CS18	B.a. 2	1.88	0.248	0.037	1.95	4.105	71.11	15.75	0.132	0.72	4.46	100.4
Barnes Gap/Crockmore	C 089254	CS20	B.a. 2	2.08	0.279	0.034	1.78	4.351	70.65	16.07	0.145	0.65	4.22	100.26
Barnes Gap/Crockmore	C 088285	CS21	B.a. 3	2.35	0.349	0.042	1.66	4.496	69.96	15.89	0.157	0.84	3.89	99.65
Barnes Gap/Crockmore	C 085252	CS24	B.a. 2	2.22	0.323	0.04	1.9	4.615	70.6	15.61	0.144	0.84	3.8	100.09
Losset	C 069217	LO01	CG	1.28	0.196	0.033	1.93	3.358	73.45	15.21	0.086	0.46	4.6	100.61
Barnes Gap/Crockmore	C 093242	CS22	CG	1.16	0.18	0.025	1.68	5.086	72.84	15.57	0.072	0.47	3.77	100.85
Barnes Gap/Crockmore	C 093242	CS22	CG	1.07	0.167	0.023	1.69	5.077	72.41	15.51	0.071	0.44	3.77	100.22
Barnes Gap/Crockmore	C 096253	CS	CG	1.69	0.202	0.039	1.67	4.628	70.77	16.55	0.089	0.54	4.2	100.37
Barnes Gap/Crockmore	C 102259	CS02	CG	1.62	0.202	0.034	1.77	4.627	72.04	15.61	0.098	0.47	4.12	100.59
Barnes Gap/Crockmore	C 102259	CS02	CG	1.64	0.202	0.033	1.76	4.614	72.07	15.63	0.102	0.48	4.13	100.66
Barnes Gap/Crockmore	C 095248	CS06	CG	1.62	0.205	0.038	1.76	4.67	72.82	15.38	0.088	0.52	4.08	101.18
Barnes Gap/Crockmore	C 095253	CS08	CG	1.7	0.209	0.037	1.73	4.643	72.16	15.81	0.092	0.55	4.19	101.12
Barnes Gap/Crockmore	C 095253	CS8	CG	1.78	0.219	0.038	1.81	4.704	71.03	16.09	0.105	0.48	4.11	100.36
Croaghacormick	C 025164	K4	CG	1.95	0.278	0.035	2.1	4.401	70.58	16.06	0.112	0.55	3.98	100.05
Croaghacormick	C 026163	K6	"CG"	1.6	0.252	0.038	2.79	2.669	71.09	16.61	0.106	0.56	4.78	100.5
Losset	C 055225	LO02	DRG	1.74	0.242	0.035	1.63	4.704	72.06	15.58	0.095	0.73	4.05	100.88
Losset	C 055225	LO03	DRG	1.75	0.261	0.034	2.18	3.889	70.63	15.34	0.097	0.83	4.17	99.18
Barnes Gap/Crockmore	C 086257	BG	DRG	1.8	0.28	0.033	2	3.948	70.49	15.47	0.13	0.8	4.25	99.2
Barnes Gap/Crockmore	C 086257	BG	DRG	1.86	0.279	0.035	1.98	3.951	70.2	16.28	0.124	0.79	4.4	99.89
Barnes Gap/Crockmore	C 086259	CS23	DRG	1.88	0.289	0.036	2.03	3.772	71.17	15.22	0.125	0.86	4.2	99.61
Barnes Gap/Crockmore	C 094265	CS26	SHEET	2.09	0.275	0.032	2.29	3.943	71.66	15.68	0.119	0.73	3.4	100.21
Barnes Gap/Crockmore	C 092264	CS25	SHEET	1.87	0.257	0.033	2.73	3.232	71.7	15.85	0.113	0.71	4.53	101.03
Losset	C 063226	LO04	SHEET	1.89	0.253	0.027	1.75	4.752	70.72	15.61	0.157	0.73	3.66	99.56
Losset	C 060227	LO05	SHEET	1.97	0.3	0.035	2.42	3.807	70.77	15.41	0.118	0.8	4.14	99.76
Barnes Gap/Crockmore	C 099264	CS10	B.a. 1	2.52	0.328	0.05	2.49	3.929	68.8	16.36	0.173	0.84	4.33	99.84
Barnes Gap/Crockmore	C 091247	CS17	B.a. 1	1.58	0.21	0.037	1.55	4.634	71.98	15.31	0.11	0.55	4.35	100.31
Carbat Gap	B 911033	CG2	TONALITE	4.11	0.606	0.047	5.1	2.498	62.96	16.08	0.277	2.53	3.76	97.96

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe2O3	TiO2	MnO	CaO	K2O	SiO2	Al2O3	P2O5	MgO	Na2O	TOTAL
Carbat Gap	B 907035	CG4	HOSTTXZ	0.94	0.132	0.027	1.87	3.804	72.82	16.09	0.054	0.35	4.56	100.64
Carbat Gap	B 921044	CG6	"GM 1"	3.79	0.633	0.036	4.57	2.679	63.63	19.01	0.274	1.56	4.47	100.65
Cocks Heath Hill	C 125328	CH4	TONALITE	5.73	0.927	0.077	4	3.646	58.65	18.39	0.366	3.36	3.88	99.02
Cocks Heath Hill	C 125328	CH4	TONALITE	5.69	0.913	0.077	4.07	3.606	59.28	18.61	0.383	3.32	3.96	99.9
Cocks Heath Hill	C 125327	CH03	FINE DYKE	3.05	0.47	0.057	3.57	1.724	67.6	16.78	0.186	1.53	5.14	100.1
Cocks Heath Hill	C 125327	CH02	PINK PORP	1.39	0.213	0.023	1.81	4.806	72.39	15.21	0.084	0.5	4.02	100.43
Cocks Heath Hill	C 113308	CH06	PINK PORP	1.43	0.205	0.025	1.64	5.07	72.11	15.3	0.102	0.54	3.95	100.37
Cocks Heath Hill	C 113308	CH06	PINK PORP	1.42	0.203	0.025	1.63	5.071	71.93	15.24	0.098	0.51	3.88	100.01
Poisoned Glen	B 936167	PG04	PINK PORP	1.46	0.22	0.027	1.62	4.743	71.36	14.9	0.089	0.51	3.87	98.8
Meenderryherk	B 837092	TB10	PINK PORP	1.76	0.248	0.039	1.03	4.775	71.22	15.82	0.096	0.68	4.53	100.19
Commeen	B 920146	Z03	"SRU 3:CWBG"	1.15	0.153	0.024	1.73	4.161	74.5	15.14	0.018	0.38	4.25	101.51
Crobane Hill	MAP A	CBH02	CBH 2	1.94	0.24	0.029	2.22	4.127	70.4	15.81	0.114	0.6	3.95	99.42
Crobane Hill	MAP A	CBH02*	CBH 2	1.87	0.239	0.028	2.26	4.138	71.44	15.23	0.127	0.61	3.85	99.79
Crobane Hill	MAP A	CBH08	CBH 2	2.19	0.297	0.037	2.04	4.291	69.81	15.23	0.125	0.64	3.73	98.4
Crobane Hill	MAP A	CBH14	CBH 2	1.86	0.244	0.035	1.73	4.375	71.04	16.07	0.099	0.58	4.06	100.08
Crobane Hill	MAP A	CBH17	CBH 2	1.47	0.188	0.02	1.22	5.647	72.75	15.06	0.054	0.46	2.99	99.86
Crobane Hill	MAP A	CBH22	CBH 2	2.07	0.283	0.036	2.06	4.303	69.55	15	0.118	0.67	3.61	97.71
Crobane Hill	MAP A	CBH01	CBH 1	2.36	0.322	0.04	2.74	2.615	70.03	15.46	0.111	0.97	4.23	98.88
Crobane Hill	MAP A	CBH01*	CBH 1	2.29	0.322	0.038	2.8	2.637	69.46	14.79	0.114	1	4.05	97.5
Crobane Hill	MAP A	CBH05	CBH 1	2.33	0.288	0.036	2.79	2.665	69.65	15.39	0.112	0.79	4.21	98.26
Crobane Hill	MAP A	CBH16	CBH 1	2.47	0.331	0.044	3.11	2.131	69.11	16.41	0.126	0.97	4.68	99.38
Crobane Hill	MAP A	CBH09	CBH 5	1.92	0.233	0.031	1.33	4.852	71.08	15.48	0.096	0.47	3.8	99.29
Crobane Hill	MAP A	CBH10	CBH 5	1.66	0.224	0.028	2.12	3.513	71.14	16.17	0.081	0.58	4.24	99.76
Croaghleconnel	B 853058	GH26	"CBH 3"	0.85	0.109	0.016	0.9	5.655	75.55	13.39	0.044	0.2	3.34	100.05
Doocharry Synform	B 860073	DS02	"CBH 3"	0.79	0.091	0.014	1.17	5.359	73.18	14.76	0.02	0.16	3.81	99.35
Commeen	B 915127	Z04	"CBH 3"	1.35	0.164	0.012	1.55	5.391	72.28	14.74	0.065	0.4	3.41	99.37
Crobane Hill	MAP A	CBH04	CBH 3	0.98	0.115	0.016	1.49	4.714	73.27	14.9	0.043	0.24	3.57	99.33
Crobane Hill	MAP A	CBH04*	CBH 3	0.95	0.115	0.016	1.49	4.56	73.22	14.07	0.031	0.23	3.42	98.1
Crobane Hill	MAP A	CBH07	CBH 3	1.11	0.132	0.018	1.72	3.932	73.79	14.96	0.034	0.26	3.99	99.94
Crobane Hill	MAP A	CBH12	CBH 3	0.42	0.032	0.008	1.15	5.34	74.21	15.42	0.024	0.06	3.77	100.43
Crobane Hill	MAP A	CBH13	CBH 3	1	0.117	0.018	1.7	4.245	73.25	15.31	0.027	0.29	3.97	99.93
Crobane Hill	MAP A	CBH15	CBH 3	0.7	0.067	0.013	1.03	5.486	74.38	15.54	0.024	0.29	3.19	100.71

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe2O3	TiO2	MnO	CaO	K2O	SiO2	Al2O3	P2O5	MgO	Na2O	TOTAL
Crobane Hill	MAP A	CBH03	CBH 4*	3.06	0.498	0.029	4.38	1.747	66.49	15.91	0.179	1.66	3.48	97.34
Crobane Hill	MAP A	CBH11	CBH 4	3.05	0.466	0.042	4.57	1.518	66.94	17.46	0.199	1.32	4.35	99.91
Crobane Hill	MAP A	CBH06	CBH 4	2.83	0.447	0.037	4.16	1.504	66.51	16.3	0.169	1.46	4.08	97.51
Crobane Hill	MAP A	CBH06*	CBH 4	2.73	0.441	0.035	4.18	1.536	66.03	15.75	0.17	1.54	3.97	96.39
Crobane Hill	MAP A	CBH11	CBH 4	3	0.452	0.043	4.35	1.478	65.88	17.58	0.181	1.46	4.43	98.87
Crobane Hill	MAP A	CBH21	CBH 4	3.38	0.476	0.045	3.77	1.864	66.41	17.29	0.187	1.48	4.61	99.52
Galwollie Hill	MAP C	GH07	GWH 1	2.9	0.408	0.049	3.42	2.429	67.39	16.94	0.156	1.44	4.22	99.35
Galwollie Hill	MAP C	GH07	GWH 1	2.83	0.404	0.049	3.44	2.42	67.53	17.2	0.162	1.46	4.29	99.78
Galwollie Hill	MAP C	GH07*	GWH 1	2.82	0.406	0.048	3.45	2.446	68.7	15.9	0.156	1.45	4.08	99.46
Galwollie Hill	MAP C	GH14	GWH 1	2.76	0.391	0.059	2.47	1.686	68.7	16.33	0.153	1.24	5.26	99.06
Galwollie Hill	MAP C	GH20	GWH 1	3.06	0.448	0.058	3.34	2.166	67.29	16.21	0.166	1.57	4.26	98.57
Galwollie Hill	MAP C	GH21	GWH 1	2.62	0.353	0.065	3.21	1.906	68.58	16.86	0.145	1.17	4.92	99.82
Galwollie Hill	MAP C	GH27	GWH 1	2.44	0.349	0.044	3.62	1.473	69.54	16.63	0.212	1.17	4.89	100.36
Galwollie Hill	MAP C	GH27	GWH 1	2.39	0.342	0.043	3.57	1.441	68.58	16.21	0.207	1.13	4.86	98.79
Galwollie Hill	MAP C	GH29	GWH 5	1.09	0.128	0.03	1.2	4.989	73.68	14.69	0.037	0.23	4.03	100.11
Galwollie Hill	MAP C	GH03	GWH 5	1.26	0.13	0.035	0.82	5.156	74.49	14.43	0.061	0.28	3.68	100.34
Galwollie Hill	MAP C	GH04	GWH 5	1.09	0.122	0.033	0.81	5.376	73.91	15.19	0.039	0.23	3.92	100.72
Galwollie Hill	MAP C	GH04	GWH 5	1.07	0.122	0.034	0.81	5.394	73.89	15.34	0.037	0.22	3.94	100.86
Galwollie Hill	MAP C	GH12	GWH 5	1.24	0.146	0.042	0.93	5.111	72.95	15.31	0.054	0.38	4.12	100.29
Galwollie Hill	MAP C	GH13	GWH 5	1.13	0.128	0.031	0.63	5.176	73.24	15.79	0.051	0.32	4.15	100.65
Galwollie Hill	MAP C	GH15	GWH 5	1.3	0.162	0.046	0.93	5.022	72.4	15.46	0.057	0.37	4.12	99.88
Galwollie Hill	MAP C	GH18	GWH 5	1.07	0.123	0.027	0.47	5.492	74.46	14.78	0.047	0.34	3.7	100.5
Galwollie Hill	MAP C	GH17	GWH 5	1.43	0.169	0.043	0.94	5.16	72.73	14.83	0.071	0.37	4.05	99.8
Galwollie Hill	MAP C	GH22	GWH 5	1.12	0.121	0.034	0.71	4.901	74.28	14.88	0.049	0.26	3.98	100.33
Galwollie Hill	MAP C	GH24	GWH 5	1.35	0.166	0.05	1.09	4.874	72.62	15.15	0.074	0.29	4.12	99.79
Galwollie Hill	MAP C	GH25	GWH 3	1.63	0.194	0.034	2.03	3.429	73.7	14.8	0.077	0.4	4.56	100.84
Galwollie Hill	MAP C	GH05	GWH 3	1.59	0.189	0.047	1.33	4.461	72.07	15.82	0.11	0.46	4.1	100.18
Galwollie Hill	MAP C	GH05*	GWH 3	1.54	0.189	0.046	1.36	4.484	72.79	14.68	0.114	0.44	3.98	99.63
Galwollie Hill	MAP C	GH06	GWH 3	1.55	0.178	0.035	0.81	4.589	71.42	16.2	0.109	0.59	4.56	100.04
Galwollie Hill	MAP C	GH10	GWH 3	1.73	0.214	0.044	1.4	4.426	71.9	15.88	0.106	0.57	4.25	100.52
Galwollie Hill	MAP C	GH19	GWH 3	1.67	0.209	0.042	1.21	4.694	71.04	16.25	0.104	0.58	4.15	99.94
Galwollie Hill	MAP C	GH23	GWH 3	2.9	0.365	0.072	1	3.967	71.53	14.91	0.117	1.01	3.92	99.78

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe2O3	TiO2	MnO	CaO	K2O	SiO2	Al2O3	P2O5	MgO	Na2O	TOTAL
Galwollie Hill	MAP C	GH09	GWH 4	1.14	0.117	0.043	0.84	5.061	74.35	14.46	0.053	0.21	3.9	100.18
Galwollie Hill	MAP C	GH16	GWH 4	1.12	0.112	0.035	0.9	5.095	72.93	15.85	0.063	0.22	4.23	100.56
Galwollie Hill	MAP C	GH01	GWH 4	1.69	0.207	0.031	1.84	4.296	71.46	16.05	0.081	0.45	4.2	100.3
Galwollie Hill	MAP C	DS04	GWH 4	2.25	0.292	0.032	1.77	4.395	70.58	15.4	0.149	0.67	4.31	99.85
Galwollie Hill	MAP C	GH31	GWH 4	1.75	0.237	0.029	1.83	4.813	70.87	15.26	0.124	0.52	3.88	99.31
Meensnee Hill (Bullaba)	C 990156	MH02	"GM 4"	1.88	0.281	0.031	1.86	5.076	69.8	15.55	0.12	0.91	3.52	99.03
Glendowan Mountains	MAP I	CU05	GM 3	2.18	0.313	0.029	2.43	4.157	70.68	16.11	0.139	0.8	3.98	100.81
Glendowan Mountains	C 980130	CU13	GM 3	1.7	0.248	0.028	2.56	3.305	71.51	15.8	0.1	0.75	4.24	100.24
Glendowan Mountains	C 981131	CU16	GM 3	1.43	0.207	0.022	2.37	3.56	72.42	15.29	0.07	0.59	4.07	100.03
Glendowan Mountains	MAP I	CU06	GM 3	1.77	0.244	0.022	1.81	5.462	71.86	15.49	0.133	0.57	3.37	100.72
Glendowan Mountains	MAP II	CU20	GM 3	2.29	0.324	0.025	1.98	3.53	71.27	15.4	0.036	0.83	4.23	99.92
Glendowan Mountains	MAP I	CU01	GM 3	2.66	0.354	0.038	3.55	1.281	70.8	16.26	0.142	1.04	5.11	101.25
Glendowan Mountains	MAP I	CU08	GM 3	2.57	0.367	0.035	3.07	2.426	70.52	16.12	0.143	1.05	4.59	100.89
Glendowan Mountains	MAP II	CU21	GM 3	2.92	0.441	0.032	2.91	1.664	69.49	16.24	0.036	1.1	5.22	100.05
Glendowan Mountains	MAP II	CU23	GM 3	2.91	0.422	0.033	2.74	1.882	70.06	16.05	0.026	1.1	4.99	100.2
Glendowan Mountains	MAP I	CU04	GM 3	2.71	0.468	0.036	3.32	2.601	69.06	16.18	0.18	1.17	4.16	99.88
Glendowan Mountains	MAP I	CU10	GM 1	2.99	0.456	0.042	3.25	2.666	67.91	15.56	0.199	1.2	3.92	98.19
Glendowan Mountains	MAP I	CU10	GM 1	3.07	0.474	0.044	3.26	2.704	67.55	15.95	0.199	1.23	4.02	98.51
Glendowan Mountains	C 980130	CU11	GM 1	2.42	0.335	0.03	3.22	2.576	70.85	15.4	0.122	1.01	4.11	100.07
Glendowan Mountains	C 978129	CU17	GM 1	2.75	0.391	0.041	3.27	2.017	69.57	15.74	0.134	1.18	4.36	99.45
Glendowan Mountains	B 950125	CU27	GM 1	3.27	0.5	0.038	3.91	1.622	67.01	15.45	0.196	1.51	3.7	97.26
Carbat Gap	B 935067	CG7	"GM 1"	3.03	0.541	0.034	4.65	1.846	66.92	17.83	0.238	1.19	4.22	100.5
Glendowan Mountains	B 951134	CU29	GM 4	2.21	0.313	0.04	2.59	3.212	70.42	16	0.203	0.83	4.42	100.24
Glendowan Mountains	MAP I	CU02	GM 4	1.59	0.236	0.021	1.96	4.668	72.45	15.71	0.101	0.54	3.92	101.21
Glendowan Mountains	B 951135	CU26	GM 4	2.1	0.347	0.029	3.3	2.672	68.99	16.56	0.138	1.18	4.25	99.58
Glendowan Mountains	MAP I	CU07	GM 4	1.7	0.228	0.026	1.88	5.032	71.73	15.97	0.137	0.62	3.82	101.23
Glendowan Mountains	C 965135	CU18	GM 6	1.59	0.222	0.023	1.89	3.6	73.65	15.27	0.034	0.55	4.3	101.13
Glendowan Mountains	C 965135	CU18	GM 6	1.59	0.223	0.024	1.88	3.622	73.24	15.26	0.036	0.56	4.27	100.7
Glendowan Mountains	MAP II	CU19	GM 6	1.26	0.178	0.018	1.23	4.745	73.22	14.86	0.041	0.38	3.65	99.58
Glendowan Mountains	MAP II	CU19	GM 6	1.26	0.18	0.018	1.28	4.934	73.64	14.83	0.034	0.39	3.84	100.41
Glendowan Mountains	B 957127	CU24	GM 6	1.39	0.201	0.016	1.52	4.547	72.31	15.35	0.036	0.49	4.04	99.9
Glendowan Mountains	B 956130	CU25	GM 6	1.23	0.165	0.01	1.63	4.263	72.67	15.22	0.047	0.48	3.91	99.62

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe2O3	TiO2	MnO	CaO	K2O	SiO2	Al2O3	P2O5	MgO	Na2O	TOTAL
Glendowan Mountains	C 951135	CU30	GM 5	1.07	0.137	0.017	1.6	4.735	72.62	15.05	0.047	0.47	3.8	99.55
Glendowan Mountains	C 952133	CU31	GM 5	1.09	0.139	0.018	1.63	4.089	73.62	14.76	0.052	0.34	4.09	99.84
Glendowan Mountains	MAP I	CU03	GM 5	1	0.148	0.016	1.37	6.393	72.63	15.63	0.064	0.31	3.18	100.72
Glendowan Mountains	C 951135	CU30	GM 5	1.05	0.133	0.017	1.56	4.612	72.61	14.86	0.049	0.45	3.69	99.03
Leahanmore	C 017167	K03	"GM 5"	1.17	0.184	0.024	2.06	3.428	73.31	14.62	0.024	0.47	4.18	99.46
Leahanmore	C 019168	K03-B	"GM 5"	0.98	0.153	0.02	2.01	3.485	74.82	14.88	0.024	0.38	4.34	101.08
Lackagh Bridge	C 096310	LB06	FINE DYKE	2.52	0.418	0.046	3.08	2.997	69.38	15.77	0.154	1.17	4.26	99.79
Lackagh Bridge	C 096310	LB07	FINE DYKE	2.2	0.299	0.04	3.2	2.784	71.01	15.09	0.114	0.89	3.99	99.61
Lackagh Bridge	C 096310	LB07	FINE DYKE	2.18	0.294	0.04	3.13	2.735	70.24	14.94	0.11	0.89	3.95	98.51
Meensnee Hill	C990156	MH1	"GM 2"	3.58	0.532	0.054	3.91	1.774	66.38	17.72	0.203	1.22	5.01	100.37
Kinneveagh	C 012182	K1	"GM 2"	2.37	0.371	0.03	2.95	3.451	70.16	16.11	0.175	0.79	3.84	100.25
Kinneveagh	C 012182	K2	"GM 1"	2.7	0.426	0.034	3.27	2.97	69.34	16.78	0.178	0.87	4	100.57
Kinneveagh	C 012182	K2	"GM 1"	2.76	0.466	0.035	3.11	3.095	68.82	15.84	0.181	1.06	3.74	99.11
Sruhanavarnis	MAP B	SA255	SRU 3: CWBG	1.27	0.17	0.02	1.87	4.065	73.19	14.83	0.059	0.41	4.08	99.97
Sruhanavarnis	MAP B	SA256	SRU 3: CWBG	1.51	0.199	0.026	2.23	3.31	73.19	14.68	0.067	0.45	4.42	100.07
Sruhanavarnis	MAP B	SA006	SRU 1	3.44	0.49	0.061	2.45	3.563	66.22	17.31	0.249	1.13	4.4	99.32
Sruhanavarnis	MAP B	SA006*	SRU 1	3.3	0.487	0.057	2.48	3.574	66.37	16.88	0.249	1.14	4.3	98.83
Sruhanavarnis	MAP B	SA009	SRU 1	3.79	0.588	0.046	3	3.074	65.71	16.96	0.254	1.35	4.27	99.04
Sruhanavarnis	MAP B	SA020	SRU 1	3.74	0.501	0.037	2.49	4.285	65.48	17.12	0.231	1.24	3.99	99.1
Sruhanavarnis	MAP B	SA122	SRU 1	3.76	0.466	0.054	3.35	1.851	66.79	16.73	0.182	1.45	4.85	99.47
Sruhanavarnis	SUB-MAP9	SA223	SRU 1	3.89	0.541	0.041	3.59	1.954	64.97	17.5	0.225	1.45	4.99	99.15
Sruhanavarnis	SUB-MAP9	SA232	SRU 1	3.59	0.596	0.036	3.27	1.917	65.71	17.15	0.118	1.32	4.81	98.53
Sruhanavarnis	MAP B	SA250	SRU 1	3.29	0.563	0.044	2.96	3.394	66.33	16.11	0.321	1.17	4.33	98.53
Sruhanavarnis	MAP B	SA251	SRU 1	3.39	0.437	0.04	3.44	1.453	67.74	16.36	0.192	1.08	5.28	99.4
Sruhanavarnis	MAP B	SA074	SRU 1	2.65	0.392	0.034	2.35	3.081	69.96	16.29	0.103	0.97	4.32	100.15
Sruhanavarnis	MAP B	SA125	SRU 1	3.71	0.443	0.056	3.23	1.749	67.16	16.65	0.18	1.52	5	99.7
Sruhanavarnis	MAP B	SA125	SRU 1	3.82	0.463	0.059	3.21	1.796	66.96	16.68	0.177	1.54	5.01	99.71
Sruhanavarnis	SUB-MAP10	SA245	DMG DYKE	3.04	0.435	0.033	2.17	4.687	68.01	15.6	0.196	0.84	3.71	98.72
Sruhanavarnis	SUB-MAP10	SA248	"SRU 1"	2.24	0.325	0.027	2.46	2.345	72.98	14.83	0.027	0.78	4.55	100.56
Sruhanavarnis	MAP B	SA017	LATE DYKE	0.83	0.1	0.023	1.07	4.071	73.09	16.29	0.032	0.18	4.69	100.37
Sruhanavarnis	MAP B	SA004	LATE DYKE	1.51	0.179	0.028	0.93	5.063	71.98	16.3	0.065	0.37	4.05	100.48
Sruhanavarnis	MAP B	SA007	LATE DYKE	1.8	0.2	0.035	1.48	5.006	71.26	16.44	0.087	0.48	3.98	100.77

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe2O3	TiO2	MnO	CaO	K2O	SiO2	Al2O3	P2O5	MgO	Na2O	TOTAL
Sruhanavarnis	MAP B	SA018	LATE DYKE	1.41	0.204	0.019	1.47	4.962	71.14	16.16	0.03	0.58	3.74	99.72
Sruhanavarnis	MAP B	SA029	LATE DYKE	1.36	0.204	0.014	1.77	4.469	71.47	16.35	0.096	0.41	4.07	100.21
Sruhanavarnis	MAP B	SA202	LATE DYKE	2.78	0.371	0.031	2.82	2.793	69.89	16.37	0.167	0.86	4.02	100.1
Sruhanavarnis	MAP B	SA217	LATE DYKE	2.68	0.386	0.034	2.62	2.856	69.29	16.53	0.175	0.94	4.37	99.88
Sruhanavarnis	MAP B	SA218	LATE DYKE	2.12	0.282	0.038	2.19	3.831	69.87	16.76	0.115	0.7	4.08	99.99
Sruhanavarnis	MAP B	SA094	SRU 2	2.07	0.306	0.026	2.37	2.395	72.09	15.49	0.099	0.85	4.45	100.14
Sruhanavarnis	MAP B	SA023*	SRU 2	1.83	0.281	0.023	2.51	1.499	74.8	14.29	0.102	0.66	4.55	100.55
Sruhanavarnis	MAP B	SA023*	SRU 2	1.81	0.274	0.024	2.52	1.478	75.13	14.24	0.106	0.65	4.6	100.85
Sruhanavarnis	MAP B	SA003*	SRU 2	2.2	0.332	0.026	2.9	1.367	73.57	14.36	0.049	0.81	4.44	100.06
Sruhanavarnis	MAP B	SA062	SRU 2	2.2	0.313	0.029	2.77	1.227	72.84	14.87	0.094	0.79	4.58	99.72
Sruhanavarnis	MAP B	SA003	SRU 2	2.28	0.329	0.027	2.85	1.341	72.4	15.35	0.057	0.86	4.55	100.05
Sruhanavarnis	MAP B	SA086	SRU 2	2.34	0.363	0.028	2.63	1.297	71.71	15.73	0.077	1.04	4.93	100.15
Sruhanavarnis	MAP B	SA165	SRU 2	2.4	0.305	0.035	2.21	3.392	70.27	16.35	0.094	0.88	4.45	100.39
Sruhanavarnis	MAP B	SA174	SRU 2	2.6	0.324	0.035	2.31	2.503	69.44	16.96	0.122	0.86	5	100.16
Sruhanavarnis	MAP B	SA106	SRU 2	2.62	0.373	0.037	2.55	2.796	68.84	16.38	0.133	1.08	4.63	99.43
Sruhanavarnis	MAP B	SA131	SRU 2	2.86	0.381	0.045	2.63	2.688	69.03	16.32	0.119	0.97	4.68	99.72
Sruhanavarnis	MAP B	SA019	SRU 2	2.9	0.413	0.034	2.95	1.504	68.25	16.95	0.064	1.12	5.07	99.25
Sruhanavarnis	MAP B	SA071	SRU 2	2.9	0.418	0.036	3.08	2.165	67.41	17.03	0.168	1.09	4.89	99.19
Sruhanavarnis	MAP B	SA117	SRU 2	2.93	0.39	0.046	2.7	2.777	67.99	16.53	0.146	1.26	4.57	99.35
Sruhanavarnis	MAP B	SA153	SRU 2	3.01	0.433	0.045	3.12	2.101	67	16.92	0.17	1.39	5.03	99.23
Sruhanavarnis	MAP B	SA008	SRU 2	3.12	0.42	0.032	2.58	1.607	71.29	15.46	0.168	1.13	4.43	100.24
Sruhanavarnis	MAP B	SA179	SRU 2	3.16	0.443	0.048	2.69	2.926	66.86	16.98	0.17	1.3	4.57	99.14
Sruhanavarnis	MAP B	SA177*	SRU 2	3.24	0.417	0.05	3.3	1.679	67.59	15.68	0.165	1.36	4.84	98.32
Sruhanavarnis	MAP B	SA008	SRU 2	3.26	0.447	0.03	2.55	1.714	70.96	15.56	0.167	1.18	4.4	100.28
Sruhanavarnis	MAP B	SA183	SRU 2	3.31	0.457	0.051	3.16	1.967	66.95	16.93	0.169	1.35	5.03	99.38
Sruhanavarnis	MAP B	SA177	SRU 2	3.34	0.421	0.052	3.27	1.679	67.52	16.5	0.162	1.37	4.97	99.3
Sruhanavarnis	MAP B	SA140	SRU 2	3.39	0.421	0.054	3.28	1.928	66.87	16.75	0.173	1.45	4.93	99.25
Sruhanavarnis	MAP B	SA042	SRU 2	3.5	0.524	0.042	3.36	1.962	65.75	17.69	0.048	1.34	4.98	99.2
Sruhanavarnis	MAP B	SA063	SRU 2	3.52	0.528	0.044	2.91	1.79	67.64	16.32	0.037	1.27	4.76	98.82
Sruhanavarnis	MAP B	SA065	SRU 2	3.54	0.537	0.04	2.69	1.93	67.97	16.98	0.024	1.07	4.94	99.71
Sruhanavarnis	MAP B	SA115	SRU 3 : PPG	1.45	0.21	0.025	1.41	5.095	72.17	15.78	0.077	0.58	3.62	100.41
Sruhanavarnis	SUB-MAP10	SA247	SRU 3: PPBG	1.7	0.237	0.024	1.78	4.847	70.48	15.94	0.106	0.58	3.9	99.59

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe_2O_3	TiO_2	MnO	CaO	K_2O	SiO_2	Al_2O_3	P_2O_5	MgO	Na_2O	TOTAL
Sruhanavarnis	MAP B	SA095	SRU 3: PPBG	1.5	0.22	0.018	1.4	5.609	70.74	16.41	0.103	0.54	3.62	100.16
Sruhanavarnis	MAP B	SA095	SRU 3: PPBG	1.53	0.227	0.019	1.42	5.618	70.61	16.33	0.106	0.57	3.59	100.01
Sruhanavarnis	MAP B	SA160	SRU 3: PPBG	1.83	0.214	0.025	1.84	5.394	69.84	16.97	0.116	0.59	3.86	100.68
Sruhanavarnis	MAP B	SA121	SRU 3: PPBG	2	0.266	0.03	1.8	5.166	69.04	16.4	0.131	0.75	3.82	99.4
Sruhanavarnis	MAP B	SA158	SRU 3: PPBG	2.08	0.262	0.034	1.9	3.904	71.7	15.06	0.109	0.71	3.92	99.68
Sruhanavarnis	MAP B	SA211	SRU 3: PPBG	2.08	0.3	0.028	1.87	5.232	68.6	16.95	0.148	0.7	3.92	99.82
Sruhanavarnis	MAP B	SA175	SRU 3: PPBG	2.12	0.275	0.025	1.92	4.873	69.43	16.74	0.13	0.7	4.01	100.23
Sruhanavarnis	MAP B	SA126	SRU 3: PPBG	2.15	0.295	0.044	1.87	4.923	68.85	16.41	0.134	0.85	3.83	99.37
Sruhanavarnis	MAP B	SA193	SRU 3: PPBG	2.18	0.316	0.035	1.94	4.739	68.95	16.91	0.139	0.79	4.1	100.09
Sruhanavarnis	MAP B	SA182	SRU 3: PPBG	2.25	0.327	0.034	1.85	4.879	68.56	16.82	0.134	0.83	4.11	99.79
Sruhanavarnis	MAP B	SA150	SRU 3:PSBGp	2.17	0.313	0.029	2.01	4.577	69.05	16.89	0.145	0.82	4.22	100.24
Sruhanavarnis	MAP B	SA107	SRU 3:PSBGp	1.88	0.273	0.025	1.88	4.165	70.28	16.29	0.118	0.78	4.2	99.91
Sruhanavarnis	MAP B	SA024*	SRU 3: PSBG	1.3	0.187	0.014	1.59	3.902	73.48	14.3	0.033	0.39	3.81	99.01
Sruhanavarnis	MAP B	SA024	SRU 3: PSBG	1.36	0.186	0.015	1.56	3.921	73.63	15.4	0.032	0.41	3.97	100.48
Sruhanavarnis	MAP B	SA076	SRU 3: PSBG	1.62	0.233	0.021	2.18	2.431	74.16	15.1	0.024	0.59	4.31	100.67
Sruhanavarnis	MAP B	SA038	SRU 3: PSBG	1.78	0.254	0.02	1.65	4.914	70.73	16.18	0.074	0.66	3.68	99.95
Sruhanavarnis	MAP B	SA073	SRU 3: PSBG	1.78	0.263	0.024	2.49	2.019	73.54	15.05	0.071	0.68	4.37	100.28
Sruhanavarnis	MAP B	SA038	SRU 3: PSBG	1.79	0.256	0.021	1.66	4.928	70.53	16.1	0.079	0.66	3.65	99.68
Sruhanavarnis	MAP B	SA031	SRU 3: PSBG	1.81	0.249	0.02	1.81	4.808	70.49	16.37	0.071	0.55	3.94	100.12
Sruhanavarnis	MAP B	SA088	SRU 3: PSBG	1.9	0.283	0.021	1.8	4.693	70.36	16.6	0.105	0.68	3.95	100.4
Sruhanavarnis	MAP B	SA088	SRU 3: PSBG	1.91	0.284	0.023	1.81	4.672	70.33	16.6	0.105	0.71	3.91	100.35
Sruhanavarnis	MAP B	SA005*	SRU 3: PSBG	2.07	0.3	0.024	1.94	4.687	68.53	16.03	0.152	0.82	3.74	98.28
Sruhanavarnis	MAP B	SA096	SRU 3: PSBG	2.09	0.306	0.024	1.8	4.476	69.68	16.38	0.102	0.77	3.89	99.52
Sruhanavarnis	MAP B	SA005	SRU 3: PSBG	2.12	0.301	0.024	1.93	4.656	69.11	16.81	0.145	0.8	3.94	99.84
Sruhanavarnis	MAP B	SA025	SRU 3: PSBG	2.26	0.28	0.027	1.8	4.364	70.64	16.17	0.134	0.69	3.99	100.34
Sruhanavarnis	MAP B	SA025	SRU 3: PSBG	2.33	0.295	0.029	1.79	4.423	70.4	16.53	0.129	0.72	4	100.64
Sruhanavarnis	MAP B	SA080	SRU 3: PSBG	2.44	0.356	0.03	2.3	4.02	68.69	16.99	0.092	0.9	4.2	100.03
Sruhanavarnis	MAP B	SA014	SRU 3:VCPG	1.55	0.218	0.02	1.58	4.651	72.21	15.86	0.026	0.57	3.79	100.47
Sruhanavarnis	MAP B	SA014	SRU 3:VCPG	1.65	0.23	0.019	1.57	4.659	71.96	15.72	0.035	0.6	3.71	100.16
Sruhanavarnis	MAP B	SA016	SRU 3:VCPG	1.32	0.187	0.017	1.44	4.865	72.41	15.97	0.032	0.55	3.78	100.56
Sruhanavarnis	MAP B	SA021	SRU 3:VCPG	1.67	0.224	0.018	1.85	5.708	69.27	16.97	0.13	0.54	3.63	100.01
Sruhanavarnis	MAP B	SA083	SRU 3:VCPG	2.01	0.3	0.024	1.61	4.866	70.91	15.34	0.111	0.75	3.49	99.4
Sruhanavarnis	SUB-MAP9	SA233	SRU 3:VCPG	1.87	0.306	0.015	1.62	4.752	70.8	16.38	0.082	0.58	3.75	100.16

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Fe2O3	TiO2	MnO	CaO	K2O	SiO2	Al2O3	P2O5	MgO	Na2O	TOTAL
Sruhanavarnis	MAP B	SA222	SRU 3:VCPG	1.27	0.196	0.014	1.66	4.505	71.71	16.33	0.067	0.51	4.02	100.29
Sruhanavarnis	MAP B	SA237	SRU 3:VCPG	1.36	0.205	0.014	1.75	4.527	71.51	16.32	0.101	0.4	4.05	100.22
Trawenagh Bay	B 819023	TB2	"GWH 1"	3.44	0.459	0.043	3.17	3.484	66.7	17.11	0.225	1.14	4.15	99.91
Trawenagh Bay	B 788027	TB5	G I "GWH 3"	2.18	0.297	0.038	1.63	4.732	70.66	15.48	0.135	0.73	3.88	99.77
Trawenagh Bay	B 7879788	TB6	G II	1.31	0.183	0.042	1.24	4.048	73.13	15.59	0.082	0.43	4.315	100.39
Trawenagh Bay (Meen)	B 834095	TB14	G I	1.7	0.232	0.034	2.06	3.919	70.83	16.64	0.103	0.63	4.36	100.51
Trawenagh Bay	B 907014	TB3	G I "GWH3"	2.49	0.332	0.041	2.25	4.486	69.13	16.48	0.153	0.92	4.11	100.39

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Nb	Zr	Y	Sr	U	Th	Pb	Ga	Zn	Cu	Ni	Cr	Rb	Ba	Ce	Nd	La	Sc	V
Barnes Gap/Crockmore	C 102267	CS09	B.a. 2	9	144.1	14.5	469.8	3.8	6.6	26.3	21.6	41	4.1	5.7	11.8	115.7	1439	56.5	16.3	23.9	3.9	18.6
Barnes Gap/Crockmore	C 099265	CS12	B.a. 3	9.5	129.4	14.4	433	2.9	3.6	26.4	19.6	41.7	5.2	7.5	15.5	113.6	1315	49.4	19.6	26.4	2.8	15
Barnes Gap/Crockmore	C 096261	CS13	B.a. 3	10.3	116.5	14.4	416.5	1.8	6	25.5	17.1	39.2	10.6	5.3	17.6	124.3	1192	47.9	13.7	24.8	4.3	10.2
Barnes Gap/Crockmore	C 095254	CS14	B.a. 2	12.7	108.7	15.7	381.1	2.5	7	25.7	19.8	37.8	6.5	4.9	10.3	134.6	1060	43.8	9.8	20.4	3.6	12.4
Barnes Gap/Crockmore	C 094261	CS16	B.a. 2	10.6	125.9	16	447.8	3	5.5	21.3	17.7	46.2	8.2	4.1	24.8	113.4	1246	68.2	23.8	38.6	3.8	17.1
Barnes Gap/Crockmore	C 089282	CS18	B.a. 2	10.1	124.3	13.8	447.6	2.8	7.8	25.2	17.4	36.7	9.8	5.8	7.3	126.2	1183	46.5	18.4	23.1	2.1	16.3
Barnes Gap/Crockmore	C 089254	CS20	B.a. 2	9.7	139.7	13.9	455	2	3.2	23.1	14.2	42.6	6.2	5.1	18.6	142.5	1409	39.3	16.1	22.4	3.6	16.8
Barnes Gap/Crockmore	C 088285	CS21	B.a. 3	10.6	159.1	16.3	490	1.7	5.9	18.9	17.4	47.4	7.5	6.4	28.2	114.9	1474	93	41.3	66.6	1.9	20.7
Barnes Gap/Crockmore	C 085252	CS24	B.a. 2	9.9	152.8	12.9	480.8	2.4	6.4	26.2	22.8	48.1	9.2	4.9	29	130.7	1548	49.4	18.8	22.4	3.2	19.9
Losset	C 069217	LO01	CG	9.4	70.8	11.8	302.2	1.7	6.1	23	17.2	38.7	2.9	3.1	8.2	129.4	654.3	17.6	9.6	12	1.2	9.9
Barnes Gap/Crockmore	C 093242	CS22	CG	5.8	60.2	9.7	337.9	1.6	0.8	23.7	15.8	28.2	5.2	4.1	11.8	145	1111	16	2.3	5.9	3.8	7.2
Barnes Gap/Crockmore	C 093242	CS22	CG	6.1	59.7	9.4	337.4	1.8	2.3	25.3	13.3	23.9	4.8	4.2	3.4	143.4	1124	11.8	1.8	5.2	2.9	8.4
Barnes Gap/Crockmore	C 096253	CS	CG	8.1	122.2	9.3	380.7		10.5	27.8	19.5	32.2	7.7	9.7	8.7	137.7	1369	35.1	18.2	21.1	4.8	18
Barnes Gap/Crockmore	C 102259	CS02	CG	9.2	108.9	14.2	354	1.2	9.3	27.1	17	31.3	5.6	5	5.5	136.3	1248	44.6	19.8	33.2	2.8	9.8
Barnes Gap/Crockmore	C 102259	CS02	CG	9.3	109.4	14.2	351.3	3.1	7.2	25.3	15.5	26	3.9	4.4	16.2	135.7	1237	46.4	16.8	32.2	3.9	13.3
Barnes Gap/Crockmore	C 095248	CS06	CG	9.9	108.1	15	344.2	2.7	8	27.3	17.6	28.3	6	7.8	3.9	143.6	1216	45.5	13.9	26.8	2.8	11.4
Barnes Gap/Crockmore	C 095253	CS08	CG	9.2	114.8	14.7	373.8	2.4	9.1	27.1	19.7	34.9	5.6	5.4	22.8	136.7	1381	50.1	15.3	25.5	3.9	12.4
Barnes Gap/Crockmore	C 095253	CS8	CG	11.1	61.5	10.9	607.8	4.6	5.2	22.9	19.1	31.3	4.1	2.2	21.4	143.5	830.9	23.5	2.7	7.9	5.7	6.8
Croaghacormick	C 025164	K4	CG	8.1	113.8	7.8	419.5	1.3	34.8	18.8	17.6	50.6	3.8	2.8	14	111.2	1575	61.9	19.7	37	4.3	18.7
Croaghacormick	C 026163	K6	"CG"	8.6	126.7	7.6	628.5	2.5	32.7	10.6	21.8	57.4	30.1	2.4	25.6	82.3	406.7	42.9	11.8	20.6	6.1	41
Losset	C 055225	LO02	DRG	9.6	122.6	14.1	360.2	3.3	8.2	28.4	15.7	38.3	4.3	4.9	29.2	140	1179	49.7	14.5	24.6	3.3	11.8
Losset	C 055225	LO03	DRG	9.3	82.3	12.6	382.5	3.6	4.1	20.9	17.9	40.6	2.2	3.2	15.1	118.7	752.7	25.8	9.9	14.5	2.7	21.7
Barnes Gap/Crockmore	C 086257	BG	DRG	9.3	110.9	13.1	426.8	3.1	5.1	20.5	17.1	41.1	6.7	5.2	18.5	110.4	923.3	40.4	13.5	24.4	3.3	13.5
Barnes Gap/Crockmore	C 086257	BG	DRG	8.1	126.5	9	430.5		8.3	18.7	19.3	39	9.3	10.6	21.1	111.6	958.3	37.3	14.7	25.6	3.5	22.9
Barnes Gap/Crockmore	C 086259	CS23	DRG	10.3	118.2	13.7	439.3	3.7	6	21	17.8	42.8	7.4	4.7	18.3	109.9	925.9	47.1	15.4	26.7	5.2	17.8
Barnes Gap/Crockmore	C 094265	CS26	SHEET	7.6	153.8	13.8	477.9	3.7	8.6	26	15.8	36.7	5.3	7.4	17.1	97.9	1801	53.4	17.6	31.9	3.4	15.8
Barnes Gap/Crockmore	C 092264	CS25	SHEET	8.5	105.5	14.9	450.9	2.9	5.3	20.7	17.7	37.9	3.3	4.5	23.8	116	1068	39.1	14.4	19	4.4	13.6
Losset	C 063226	LO04	SHEET	9	123.7	15.6	446.5	3.4	8.6	29.5	14.9	36.2	4.1	8.7	7.5	126.4	1693	59.3	20.5	28.5	5.4	12.3
Losset	C 060227	LO05	SHEET	-8.4	134.9	12.9	454.3	4.6	9.6	21	17.6	40.1	5.2	5.9	11.5	110.6	1261	48.9	20.7	33.7	3.2	22.4
Barnes Gap/Crockmore	C 099264	CS10	B.a. 1	13.1	171.4	21.1	555.6	2.1	5.1	23.6	17.1	48.6	3.9	5.7	19.1	116.9	1634	66.3	19.6	29.4	4.2	18.9
Barnes Gap/Crockmore	C 091247	CS17	B.a. 1	11.3	100	14.4	349	1.7	7.8	27	17.3	31.3	4.7	6.1	8.7	142.2	994	34.9	11.4	20.5	4.6	12.6
Carbat Gap	B 911033	CG2	TONALITE	8	91.6	14.7	668.7	4.7	3.6	14.4	22.1	61.4	28.3	10.1	42.2	69.9	480.6	57	20.4	28.8	9.5	109.4
Carbat Gap	B 907035	CG4	HOSTTXZ	8.7	158.8	7.9	817.8	1.9	33.7	11.5	25.5	55.9	57.2	7.4	20.3	89.9	708.7	52.9	12	30.4	4.1	75.6
Carbat Gap	B 921044	CG6	"GM 1"	7.1	103	6.7	834.1	1.9	28.9	10.7	23.5	61.2	5.4	0	15.9	49.1	561.4	71	37.8	64.2	3.6	53.9
Cocks Heath Hill	C 125328	CH4	TONALITE	8.9	92.3	10.8	433.1	4.7	7.5	22.3	18.3	34.1	5.2	2.6	13.3	105.2	1070	38.3	14.3	16	2.4	12.3

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Nb	Zr	Y	Sr	U	Th	Pb	Ga	Zn	Cu	Ni	Cr	Rb	Ba	Ce	Nd	La	Sc	V
Cocks Heath Hill	C 125328	CH4	TONALITE	9.1	174.9	8.5	826.1	2.4	2.2	13.9	26.7	77.1	6.3	16.8	60.8	118.3	1261	54	11.4	26.5	5.4	92.6
Cocks Heath Hill	C 125327	CH03	FINE DYKE	15.8	133.4	12.6	524.3	1.8	8.5	17.7	21.4	64.7	7.2	2.4	20.7	91.5	224	47.1	20.1	26.3	3.9	43.8
Cocks Heath Hill	C 125327	CH02	PINK PORP	5.8	82.7	10.2	387.4	3.6	5.1	25.1	13.4	24	3.4	1.4	17.1	119.3	1250	34.9	10.9	18.5	4.2	5.9
Cocks Heath Hill	C 125327	CH02	PINK PORP	5.8	82.7	10.2	387.4	3.6	5.1	25.1	13.4	24	3.4	1.4	17.1	119.3	1250	34.9	10.9	18.5	4.2	5.9
Cocks Heath Hill	C 113308	CH06	PINK PORP	7.7	100.4	12.5	337.8	4.2	11	27	13.2	29.6	9.3	6.5	13.7	130.5	1069	39.5	14	23.6	0.7	13.2
Cocks Heath Hill	C 113308	CH06	PINK PORP	7.7	101	13.2	339.6	3.2	12.7	26.4	14.5	30.8	9.1	7.6	10.1	130.6	1061	44.3	17.3	22.1	3.7	14.5
Poisoned Glen	B 936167	PG04	PINK PORP	6.1	79.2	9.6	425.3	1.9	5	22.1	18.7	30.9	4.3	6.4	7.9	117.9	1343	46.1	17.8	23.6	3.8	11.2
Meenderryherk	B 837092	TB10	PINK PORP	10.3	77.5	7.3	448.4	2.5	34.7	15.6	25.2	36.7	2.2	0.4	7.8	118.5	728.4	24.1	8.3	12.6	3.9	14.4
Commeen	B 920146	Z03	"SRU 3:CWBG"	6.6	50.9	9.7	331.5	3	3.7	23.7	18	26.3	3.1	4.4	6.6	105.1	974.4	28.2	6.9	15.9	3.9	5.7
Crobane Hill	MAP A	CBH02	CBH 2	9.2	140	10.8	369.1		8	22.3	18.4	42.2	8.8	10.8	14.5	133.7	1487	40.9	20.3	22.2	5	21.2
Crobane Hill	MAP A	CBH02*	CBH 2	9.4	130.3	15.6	357.8	2.4	7.6	25.6	17.9	38.2	5.5	3.6	18.6	133.1	1452	34.9	17.6	27.5	5.1	12.8
Crobane Hill	MAP A	CBH08	CBH 2	9.3	132.4	8.9	345.9		4.3	22.5	19.9	44.5	6.3	9.9	13.8	128.7	1285	31.5	12	10	5.8	24.7
Crobane Hill	MAP A	CBH14	CBH 2	8.9	117.8	11.7	305.4		5.3	23.6	13.9	43.6	9.1	11.9	24.2	133.4	1257	31.5	13.2	20.9	1.9	15.8
Crobane Hill	MAP A	CBH17	CBH 2	5.4	82.3	7.3	205.1		4.9	23.7	17.6	27.4	5.2	9.6	20.5	119.9	717.8	48.7	24.8	40.2	2.4	15.4
Crobane Hill	MAP A	CBH22	CBH 2	8.8	130.5	8.9	367.4		3.9	20.9	16.5	39.6	6.3	8.7	23.3	126.2	1264	25.7	3.6	4.9	6.4	21.7
Crobane Hill	MAP A	CBH01	CBH 1	8.3	99.9	7.3	363.8		4.4	17.2	20.6	49.5	6	9.2	35.7	88.5	716.8	24.5	9.6	11.3	4.1	34.2
Crobane Hill	MAP A	CBH01*	CBH 1	9	84.5	11.4	375	1.3	2.1	18.9	19.1	46.3	5.5	4.6	42.1	88.3	639.1	19.2	11	11.4	5.4	32.6
Crobane Hill	MAP A	CBH05	CBH 1	9.8	106.5	9.9	338.4		6.4	21.4	21.2	55.1	5.8	13.5	26.5	101.8	610.3	24.8	9.4	14.1	6.6	27.2
Crobane Hill	MAP A	CBH16	CBH 1	9.6	122.5	8.8	373.8		4.7	21.3	23.9	50.7	20.9	13	19.9	90.6	417.9	37.1	12.9	14.7	7.5	33
Crobane Hill	MAP A	CBH09	CBH 5	7.7	129.5	7.7	433.1		7.5	21.2	14.9	41.4	5.2	7.6	24.9	105.8	1509	78.4	20.4	51.2	2.1	17.1
Crobane Hill	MAP A	CBH10	CBH 5	5.3	108.9	6.1	266.7		9.2	22.1	20.8	39	16.4	11.1	31.6	88.4	1012	34.6	14.3	18.6	4.8	23.8
Galwollie Hill	B 853058	GH26	"CBH 3"	8.2	70	16.4	235.1	3.4	8.3	28.6	15.7	20	7.2	6.1	17.4	156.7	871.6	31.8	12.9	18.9	1.2	2.4
Doocharry Synform	B 860073	DS02	"CBH 3"	6	48.8	9.7	243.5	3.2	5.9	26.5	15.1	18.8	2.2	5.3	7.8	139	611.2	26	8.2	11.7	0.1	1.1
Commeen	B 915127	Z04	"CBH 3"	5.9	65.9	9.5	338.6	2.9	3.4	25.4	14.2	23.9	4.8	6.4	6	138.2	1074	30.6	10.4	16.2	4.5	6.8
Crobane Hill	MAP A	CBH04	CBH 3	4.6	66.9	5.5	316.9		7	25.2	17	20.8	8.5	9.4	15.5	113.6	1541	23.9	6.4	11.1	4.3	11.9
Crobane Hill	MAP A	CBH04*	CBH 3	4.3	52.3	10	311.7	1.7	3.7	26.2	18.8	23.4	6	5.7	7.2	114.8	1460	21	8.9	13.5	1.3	4.9
Crobane Hill	MAP A	CBH07	CBH 3	4.9	74.4	5	320.7		9.1	22.7	18.3	26.2	9.5	9.3	14.5	104	1490	26.5	11.4	13.6	3.2	10.1
Crobane Hill	MAP A	CBH12	CBH 3	1.4	46.4	3.9	219.1		1.1	32.9	15.5	9.8	4.6	11.2	24.4	117.6	583.6	6.6	1.5	0.6	2.4	5.1
Crobane Hill	MAP A	CBH13	CBH 3	4.2	68.7	4.8	321.9		5	21.1	17.9	20.3	10.4	10.6	10	105.6	1531	35.7	14.9	15.8	3.3	13.7
Crobane Hill	MAP A	CBH15	CBH 3	3.2	50.5	4.8	285		6	22	17	6.7	9.8	8.6	29	120.1	1304	22.7	3.4	11.9	1.5	8.6
Crobane Hill	MAP A	CBH03	CBH 4*	2.9	148.1	3.4	595.3		5.7	8.9	17.2	51.2	18.2	14.8	27.8	59.9	610.6	40.5	17.5	28.1	7	48.9
Crobane Hill	MAP A	CBH11	CBH 4	14.6	154.1	13.2	489	3.2	2	13.2	25.1	73.6	-0.3	-2	23.9	83.7	218.6	39.7	14.9	25.1	5.7	42.8
Crobane Hill	MAP A	CBH06	CBH 4	5.2	147	7	559.7		7	10.7	17.6	52.5	20.1	15.4	23.6	71.8	570.8	39.4	14.5	29.8	9.5	49.4
Crobane Hill	MAP A	CBH06*	CBH 4	8.6	130.2	9.2	580.9	3.3	3.7	13.4	18.9	43.7	19.4	5	25.3	74.1	526.1	47.9	19.1	28.6	4.3	40.4
Crobane Hill	MAP A	CBH11	CBH 4	4.9	150.9	8	590.2		7.7	10.1	25.5	62.9	34.1	14.4	25.5	84.4	435.2	38.1	20.4	27.4	6	42.9

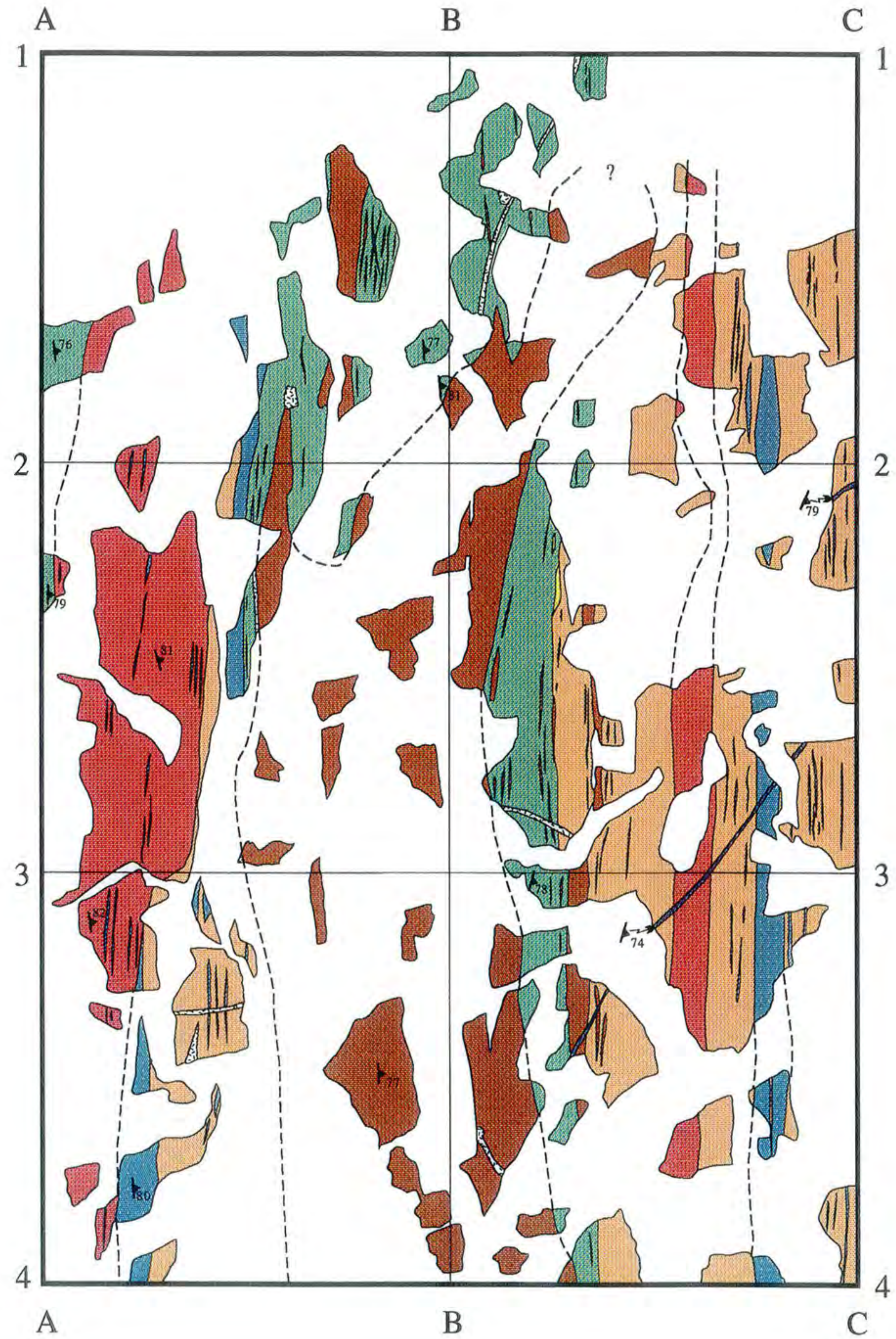
LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Nb	Zr	Y	Sr	U	Th	Pb	Ga	Zn	Cu	Ni	Cr	Rb	Ba	Ce	Nd	La	Sc	V
Crobane Hill	MAP A	CBH21	CBH 4	7.5	167.9	9.8	468.1		7.3	13.9	22.1	55.5	26.3	14	31.5	95.9	631.9	42.9	11.4	20.3	7.3	52.2
Galwollie Hill	MAP C	GH07	GWH 1	9.3	130.6	13.5	466.4		8	12.8	19.4	48.3	9.6	11.3	26.9	100.6	620.4	31.8	17.4	21.5	7.6	43.5
Galwollie Hill	MAP C	GH07	GWH 1	9.1	134.2	12.7	469.1		7	13.3	18.3	48.4	9.2	13.1	30.1	103.9	624.5	32.3	15.1	21.4	9.2	44.8
Galwollie Hill	MAP C	GH07*	GWH 1	10.7	119.6	15.2	479.8	4.7	3.7	14.7	19.9	43.4	10.6	4.2	27.2	101.9	571.7	35.2	8.1	20.4	3.9	45.1
Galwollie Hill	MAP C	GH14	GWH 1	14.8	150.4	13.5	397.8		6.4	13	25.8	66.6	6.3	8.9	25.1	114.1	304.2	34.7	9.6	10.3	9.4	34.1
Galwollie Hill	MAP C	GH20	GWH 1	11.4	140.9	13.4	469.7		7.6	14.2	22	57.3	40.8	10.3	27.7	106.7	828.3	40.1	16	21.9	6.8	49.4
Galwollie Hill	MAP C	GH21	GWH 1	13.9	124	15.9	382.9		5.4	16.7	20.6	60.5	14.4	15.4	21.5	113.4	370	32.1	15.8	17.9	7.2	32.8
Galwollie Hill	MAP C	GH27	GWH 1	11.3	62.6	12.8	568.7	5.1	1.7	15.8	22.2	59.4	14.4	2.1	20	91.2	359.9	43.3	16.6	27.7	1.7	43.9
Galwollie Hill	MAP C	GH27	GWH 1	12	61.8	12.1	572.9	5.2	4.1	16.7	23.8	59.6	15.9	1.7	32.3	91.5	376.4	56.2	15.3	25.9	9.8	43.7
Galwollie Hill	MAP C	GH29	GWH 5	16.2	71.5	33.3	208	4.4	11.3	30.7	18.5	33	5.9	3.8	17.8	189	578.2	40.8	14	21.3	2.2	3.5
Galwollie Hill	MAP C	GH03	GWH 5	13.6	105.7	10.4	176.5		8	26	22.2	38.2	5.7	8.2	7.8	174.7	754.7	41.1	14.3	14.2	0.1	9.1
Galwollie Hill	MAP C	GH04	GWH 5	11.3	75.5	12	129.1		6.7	26.9	16.1	34.1	3.7	10.5	13.2	162.3	292.3	21.7	5.9	8.1	1.4	3.8
Galwollie Hill	MAP C	GH04	GWH 5	11.1	76.3	12.7	130.5		5.9	29	17.2	29.4	3.6	9.1	14.1	160.6	287.3	19.8	8.1	9.8	3	6.1
Galwollie Hill	MAP C	GH12	GWH 5	13.9	95	14	210.1		7	25.6	17.6	33.8	7.7	10.5	20.4	175.8	793.4	23	9.5	14	3.7	6.1
Galwollie Hill	MAP C	GH13	GWH 5	12.4	84.8	12.8	177		3.5	20	15.3	26.8	7.5	8.8	12.4	184.7	743.4	20.7	9.4	13.9	2.5	8
Galwollie Hill	MAP C	GH15	GWH 5	12.9	110.6	8.6	282.7		6.4	24.3	17.5	30.5	11.4	10.2	17.1	163.4	1107	45.7	17.5	19.5	2.8	10.6
Galwollie Hill	MAP C	GH18	GWH 5	9.7	81.5	11.3	156.9		6.2	24.5	18.1	26.5	5	10.1	17.1	162.9	481.7	22.2	8.7	12.1	0.9	6.2
Galwollie Hill	MAP C	GH17	GWH 5	13.9	118.2	11.4	286.8		6.6	25.1	19	39.2	9.4	10	21.3	169.5	1133	42.3	13	19.4	0.8	10.4
Galwollie Hill	MAP C	GH22	GWH 5	16.7	84.7	17.9	111.7		4	25.6	18.1	49	6.1	8.9	17.9	196.4	295.4	15.9	2.9	4.8	0.8	7.5
Galwollie Hill	MAP C	GH24	GWH 5	13.6	110.9	12.8	269		8.2	24.1	25.5	37.2	13.2	8.9	16.9	182.7	1020	36.2	12.9	17.5	1.6	12.6
Galwollie Hill	MAP C	GH25	GWH 3	11.3	84.5	11.4	308.3	2.8	6.5	22.8	22.1	41.2	2.3	2.5	15.3	118.4	646.1	34	14.1	18.9	1.3	10.2
Galwollie Hill	MAP C	GH05	GWH 3	16.4	120.8	13.3	291.3		7.3	22.5	19.8	43.3	10.9	12.4	11.8	188	1018	35.6	9.9	7.1	3	15.4
Galwollie Hill	MAP C	GH05*	GWH 3	15.1	102.4	14.5	290.7	2.8	6.4	23.1	19.2	37.3	5.1	1.9	12.5	184.9	937.2	31.4	11	18.5	2.2	7.5
Galwollie Hill	MAP C	GH06	GWH 3	11.4	118.6	10.4	235.6		8.5	20.1	18.3	36.2	5	8.1	12.4	148.4	993.2	24.7	9.8	8.9	3.5	10.2
Galwollie Hill	MAP C	GH10	GWH 3	14.6	130.8	10.9	359.5		9.7	24	21.2	39.6	7.5	10	15.3	162.9	1143	44.3	19	23.4	4.5	15.9
Galwollie Hill	MAP C	GH19	GWH 3	12.7	122.7	10.7	320		6.8	21.3	20.7	37.1	11.2	9.5	15	156	1221	37.8	6.9	7.1	2.1	15.7
Galwollie Hill	MAP C	GH23	GWH 3	21.5	161.9	16.9	165.9		13.9	21.2	22.9	79.8	8.8	9.9	18.2	181.5	559.4	44.8	23.8	20.3	3.1	23.9
Galwollie Hill	MAP C	GH09	GWH 4	13.7	85.6	12.5	171.3		10.1	23.6	20.9	35.6	5.9	8.2	6.9	172.4	601.3	36.5	13.1	15.6	2.7	7.6
Galwollie Hill	MAP C	GH16	GWH 4	18.3	82	24.5	141.6		8.9	30.2	20.3	38.2	4.1	9.1	7.4	196.3	438.2	32.9	8.1	12.3	4.1	5.7
Galwollie Hill	MAP C	GH01	GWH 4	8	113.5	8	308.2		5.3	25.5	13.5	38.7	6.4	10.8	10	135.4	1035	26.7	3.8	9	3.5	20.8
Doocharry Synform	MAP C	DS04	GWH 4	13.1	166.4	20.8	262.4	2.4	9.6	28	20.1	52.9	1.9	2	10.2	145.5	828.1	50	18.8	24.8	5.2	21.2
Galwollie Hill	MAP C	GH31	GWH 4	9.2	114.1	18.9	328.5	2.2	5	27.2	17.2	40.3	3	3	16.3	154.1	1216	38.1	10.9	17.2	4.7	14.4
Meensnee Hill (Bullaba)	C 990156	MH02	"GM 4"	7.6	129	12.9	419.3	3.2	5.5	25.1	14.8	41.6	5.3	6.3	13.4	133.3	1761	61.1	18.8	31.7	3	19.2
Glendowan Mountains	MAP I	CU05	GM 3	5.7	171.7	6.1	497.3	1.8	6.3	18.6	18	41.7	24	3.2	8.5	99.8	1526	46	18.6	30.7	5.1	21
Glendowan Mountains	C 980130	CU13	GM 3	7.3	87.8	9.7	436.5	2.2	0.3	19.7	18.5	34.6	3.3	4.9	8.8	94.5	1086	17.8	3.6	11.6	3.2	17.1

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Nb	Zr	Y	Sr	U	Th	Pb	Ga	Zn	Cu	Ni	Cr	Rb	Ba	Ce	Nd	La	Sc	V
Glendowan Mountains	C 981131	CU16	GM 3	5.6	67.5	10.1	415.5	3.4	0.8	22.8	19	27.1	4.9	4	11.5	96.9	1396	6.3	2.7	-0.6	3.7	11.3
Glendowan Mountains	MAP I	CU06	GM 3	6.4	224.4	8.3	441.6	2.1	13.5	25.5	20.7	39.3	17.7	6.9	6.6	113.4	2240	98.8	40.7	63.7	2.9	21.4
Glendowan Mountains	MAP II	CU20	GM 3	13.1	154	15.2	307.1	3.4	11	24.6	19.3	42.2	0.5	3.5	15.1	128.3	461.1	49.2	18.4	26.6	4.4	23.5
Glendowan Mountains	MAP I	CU01	GM 3	10	119.5	8.9	392.9	2.8	6.1	15.6	18.6	56.2	2.5	2.1	15.6	72.1	151.3	51.7	12.6	24	4.9	28.5
Glendowan Mountains	MAP I	CU08	GM 3	8.7	122.7	11.9	394	1.6	5	19.4	22.4	51.4	2.4	0.4	34.7	75.7	512.9	55.8	22	31.8	4.2	32
Glendowan Mountains	MAP II	CU21	GM 3	15.7	124	14.3	347.1	2.2	10.7	20.1	23.4	57.9	1	0.2	15.6	103.7	84	49.9	18.4	27.5	3.6	35.3
Glendowan Mountains	MAP II	CU23	GM 3	15.4	131.5	12.6	333.2	2	6.4	17	23.9	55.2	1.2	3.5	29.4	104.5	123.7	38.2	13.8	19.8	2.3	35.3
Glendowan Mountains	MAP I	CU04	GM 3	7.9	147	7.9	550.3	2.5	8.5	17	17	55.9	7	3.9	15.7	71.6	990.1	94.2	34.9	63.8	6.4	48.3
Glendowan Mountains	MAP I	CU10	GM 1	11.1	158.3	14.4	489.3	2.6	11.9	20.9	22	58	29.4	5.5	17.9	78.1	713.5	91.6	28.6	51.8	4.4	55.6
Glendowan Mountains	MAP I	CU10	GM 1	9.1	160.1	14.5	469.5	1.6	5.1	21.4	24.6	58.1	27.8	5.7	19.6	79.4	730.5	97	32.4	54.8	5.3	54.4
Glendowan Mountains	C 980130	CU11	GM 1	7.9	145	9.6	479.3	3.2	6.9	18.6	20.9	47.3	7.2	3.8	10	80.8	1380	35.9	17.8	27.4	2	31.5
Glendowan Mountains	C 978129	CU17	GM 1	11.6	159.1	13.9	442.3	3.8	7.6	19.1	16.8	64.4	0.6	0.8	11.2	81.1	529.2	34.7	11	14.5	1	31.3
Glendowan Mountains	B 950125	CU27	GM 1	9.7	191.4	9.8	580.7	4.6	9	12.1	21.1	52.7	8.2	4.3	24.1	87.8	572.3	70.7	19.9	39.4	4.3	52.2
Carbat Gap	B 935067	CG7	"GM 1"	15.2	63	14.1	198.5	6.5	6.7	20.6	21.1	39.9	0.7	-0.1	11.7	149.1	544.1	18.3	11.2	9.9	1.2	9.2
Glendowan Mountains	B 951134	CU29	GM 4	11.1	115.6	16	340.7	3.3	2.4	25.1	25.7	60.1	3.6	4.4	12.1	123.5	723.3	22.3	12	14.4	5.7	27.4
Glendowan Mountains	MAP I	CU02	GM 4	7.5	110	7	297.1	3.2	6.5	26.4	18.6	41.1	5.2	3.6	8	121.8	842.1	31.3	16.7	17.5	2.6	16.4
Glendowan Mountains	B 951135	CU26	GM 4	6.7	81.2	8.6	614.5	3.6	3.3	17.2	22.9	47.2	16.8	5.2	12.6	61.5	687.6	44.4	14.1	25.6	3.7	37.3
Glendowan Mountains	MAP I	CU07	GM 4	7.3	140	7.2	375.5	3.9	14	27.8	17.3	39	3.7	3.1	5.8	118.6	1304	34.8	31.8	37.6	2.8	20.1
Glendowan Mountains	C 965135	CU18	GM 6	11.3	75.8	26.7	268.7	3.4	9	24.4	19.7	29.5	2.6	5	18.5	130.7	718.3	35.1	10.7	15.7	3.9	14.6
Glendowan Mountains	C 965135	CU18	GM 6	11.5	76.4	29.1	271.6	2.8	10	26.4	18.7	30.2	3.6	3	23.8	130.9	715.5	34	9.9	15.3	2.6	10.2
Glendowan Mountains	MAP II	CU19	GM 6	9.2	73	12.8	290.1	2.5	5.6	26.1	19.1	25.5	2.6	3.8	12.9	162.3	829.6	35.3	10.2	18.1	3.1	11.2
Glendowan Mountains	MAP II	CU19	GM 6	8.6	73.4	12.8	290.7	1.8	5.2	28.1	15.9	24.8	2.2	6.6	19.9	161.7	896.1	28	17.1	17.6	1.3	7.7
Glendowan Mountains	B 957127	CU24	GM 6	10.2	74.5	11.6	300.3	4.1	13.8	28.6	17.7	31.8	1.4	0.5	6.6	159.3	895.3	54	19.6	31.9	2.6	8.6
Glendowan Mountains	B 956130	CU25	GM 6	8.2	65.7	13.4	270.3	2.8	6.7	24	17	26.1	3.1	4.9	2	133.5	802.4	21.2	5.5	12.1	2.2	9.8
Glendowan Mountains	C 951135	CU30	GM 5	6.4	60.5	14.2	247.2	2.6	6	24.1	21.7	24.1	4	4	27.4	131.4	806.7	29.1	12.6	17	3.2	6.7
Glendowan Mountains	C 952133	CU31	GM 5	8.5	56.1	20.5	230.5	2.4	2.7	26.7	16.2	26.1	3.1	4.8	5.1	119.3	561.1	22.4	9	8.2	2.8	11.9
Glendowan Mountains	MAP I	CU03	GM 5	4.8	55.3	1.2	378.1	2	0	28	15.7	17.2	10.6	7	1.4	154.6	1660	11.3	4.3	7.1	4.1	8
Glendowan Mountains	C 951135	CU30	GM 5	6.4	62.4	12.8	246.3	2.5	7	24.4	14.3	26.1	4.8	2.5	12.8	130.6	805	23.5	8.3	12.9	2.4	2.7
Leahanmore	C 017167	K03	"GM 5"	6.8	65.4	13.2	348.7	2.7	6	23.5	15.6	28.9	5.5	3.6	3.5	101.5	774.2	26.1	6.2	11.6	2.3	7
Leahanmore	C 019168	K03-B	"GM 5"	6.8	59.7	15.5	332.6	2.6	4.7	23.3	12.3	26.9	3.9	3.9	5.7	98.5	789.1	26.8	6.3	11.6	3.3	1.6
Lackagh Bridge	C 096310	LB06	FINE DYKE	10.1	170	11.6	532.6	2.8	10.4	21.7	15.7	52.5	7.2	6.1	15.4	96.7	1105	22.5	5.6	8.9	3.9	34
Lackagh Bridge	C 096310	LB07	FINE DYKE	10.5	117.1	14.2	485.4	0.4	5.2	19	19.3	44.9	2.9	4	12.2	92.6	1076	30.4	5	10.4	4	27.9
Lackagh Bridge	C 096310	LB07	FINE DYKE	9.6	119.1	13.8	466.8	1.5	1.4	19.6	20.2	45	1.9	3.8	14.7	93.9	1083	24.7	3.5	8.8	1.6	28.4
Meensnee Hill	C990156	MH1	"GM 2"	8.9	161.7	9	827.9	1.2	1.2	11.9	26.2	77.7	8.3	17	65.2	116.4	1237	59.2	18.5	28.4	4.6	85.8
Kinneveagh	C 012182	K1	"GM 2"	7.1	162.3	8.3	716.7	1.2	31.6	17.2	22.9	62.6	13.6	3.6	18.4	89.5	2045	70.2	22.6	42.7	5.9	33.4

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Nb	Zr	Y	Sr	U	Th	Pb	Ga	Zn	Cu	Ni	Cr	Rb	Ba	Ce	Nd	La	Sc	V
Kinneveagh	C 012182	K2	"GM 1"	8.4	122.1	9.4	501.8	3.1	32.5	18	17.7	47.1	4.5	2.7	4.1	107.3	1595	43.9	15.5	30	5.2	21.5
Kinneveagh	C 012182	K2	"GM 1"	10.1	154.9	13.4	466.6	1.2	8.1	17.4	20.8	56.3	7.4	4.8	15.6	85.9	836.3	101.6	34.1	60.4	3.9	51
Sruhanavarnis	MAP B	SA255	SRU 3: CWBG	5.3	49.4	8	404.6	3.8	1.6	22.1	15.5	28.3	5	3.9	7.3	94.1	1137	17.7	1.6	5.7	4.8	4.1
Sruhanavarnis	MAP B	SA256	SRU 3: CWBG	7.2	61.8	9.5	353.2	3	1.5	22.4	20.1	30.9	3.1	3.8	8.7	91.3	820.5	21.1	8.3	8.4	5.9	18.4
Sruhanavarnis	MAP B	SA006	SRU 1	14.2	209.7	19.9	550.2		2.7	25.9	20.7	62.3	10	12.8	32.8	163.5	1666	69	31.4	34	7.6	37.9
Sruhanavarnis	MAP B	SA006*	SRU 1	13	193.9	25.3	544.5	0.9	2.2	27.1	21.5	60.6	6.2	3.5	34.7	156.8	1613	74.4	24.5	37.1	7.1	30.8
Sruhanavarnis	MAP B	SA009	SRU 1	12.8	228.5	12	656.7		6.3	17.8	21.2	75.4	18.7	10.2	20.7	128.3	1869	77	29.7	45.3	7.8	46.9
Sruhanavarnis	MAP B	SA020	SRU 1	21.2	213.8	20.9	463.7		8.5	24.9	27.7	68.1	15.3	9.6	16.9	169.8	1187	67.3	27.8	32.5	7.2	37.1
Sruhanavarnis	MAP B	SA122	SRU 1	16.2	178.2	15.1	305.7		5.1	17.8	22.2	81.4	5.4	10.3	18	111.3	111.1	27.5	20.6	27.2	9.4	23.2
Sruhanavarnis	SUB-MAP9	SA223	SRU 1	24.9	171.6	24.4	319.3		1	17.7	26.5	81.8	8.7	10.1	16.5	124.4	100.5	23.1	10.4	2.2	8.3	61.4
Sruhanavarnis	SUB-MAP9	SA232	SRU 1	18	161	23.2	354.1		25	17.9	22.2	64.6	6.2	9.8	14.3	107.5	134.1	98.6	38.3	50.1	9.7	45.5
Sruhanavarnis	MAP B	SA250	SRU 1	13.9	187.9	17	555.5	4.2	9.3	21.8	19.2	68.1	27.4	5.8	17.3	121.4	1051	56.3	15	28.8	8.2	50.8
Sruhanavarnis	MAP B	SA251	SRU 1	16	136.5	14.8	401.9	2.3	0.2	14	25.3	63.6	8.8	1.9	23	99.5	134.3	11.9	5.3	5.2	2.6	32.9
Sruhanavarnis	MAP B	SA074	SRU 1	13.9	171.9	12.3	381.6		15.4	20.6	19.5	59.1	7.7	10.9	13.3	129.4	799.1	77.9	38.4	56.5	3.5	26.9
Sruhanavarnis	MAP B	SA125	SRU 1	16.5	195.1	14.6	312.2		6.1	17.4	23.4	86.9	3.7	13.1	20.4	115.2	118.9	22.1	16.3	23.5	6.4	38.1
Sruhanavarnis	MAP B	SA125	SRU 1	16.3	188.6	14.8	309.9		5.4	17.7	25.7	84.8	4	12.6	18.9	118.1	125.9	28.7	19.2	20.8	8.7	36.5
Sruhanavarnis	SUB-MAP10	SA245	DMG DYKE	5.2	231.9	10	664.5	0.7	4.7	17.7	17.1	58.3	5.1	6.1	12.1	107.4	2725	126.2	38	63.8	3.1	31.1
Sruhanavarnis	SUB-MAP10	SA248	"SRU 1"	14.1	179.2	8	390.9	3.6	1	17.6	16.2	43.5	5.1	2.3	12.8	94.9	542	7.4	4.9	6.9	3.5	28.3
Sruhanavarnis	MAP B	SA017	LATE DYKE	4.6	52.2	3	296.7		5.2	29.4	17.8	30.1	12.6	9.1	23.8	89.7	1282	11.9	2.1	3.7	3.4	9.3
Sruhanavarnis	MAP B	SA004	LATE DYKE	11.9	131.3	8.2	308.1		3.7	25.4	18.6	30.3	8.6	9.1	20.8	127.5	1424	43.9	19.2	19.7	3.5	13.9
Sruhanavarnis	MAP B	SA007	LATE DYKE	9.5	151.9	5.9	451.6		3.7	22.6	17.5	37	12.1	10.9	17.9	111.2	1867	47.1	15.6	26.3	2.8	16.7
Sruhanavarnis	MAP B	SA018	LATE DYKE	9.5	102.8	9.1	359.8		5.3	28.7	16.4	24.1	7.1	10.2	10	146.8	1345	37.2	16.6	21	3	16.1
Sruhanavarnis	MAP B	SA029	LATE DYKE	10.5	121.7	9.3	357.5		4.1	25.6	12.7	30.8	9.6	10.9	11.1	131.3	1425	31.1	14.7	10.7	4.8	19.2
Sruhanavarnis	MAP B	SA202	LATE DYKE	5.6	196.9	7.4	365.4		18.3	16	21.5	49.1	10.1	11.1	28.7	93.7	1309	106.5	48.2	93.8	5.3	41.1
Sruhanavarnis	MAP B	SA217	LATE DYKE	9.4	170.2	14	383		13	14.3	19	55.3	14.2	12.1	35.6	112.2	1052	88.2	37.5	60.1	5.5	45.4
Sruhanavarnis	MAP B	SA218	LATE DYKE	7.4	122.3	11.5	287.2		12.5	20.7	21.4	49.7	7.2	10.7	25	119.2	777.5	52.5	10.3	4.2	4.3	27.3
Sruhanavarnis	MAP B	SA094	SRU 2	7.6	120.2	9.5	442.8		7.6	16.6	18	39.8	4.4	10.1	20.6	87.2	608	53.4	27	32	5.4	25.8
Sruhanavarnis	MAP B	SA023*	SRU 2	12.9	85.3	18.4	377	2.2	8.6	14.6	15.2	33.9	1.1	-0.4	18.6	68.2	150.5	51.5	17.8	29.7	4.5	9.3
Sruhanavarnis	MAP B	SA023*	SRU 2	13.1	88.1	19.2	377.9	3.5	10.5	16.4	16.5	30.6	3.5	0.5	16.5	67.4	151.9	54.5	18.6	31.2	2.9	10.2
Sruhanavarnis	MAP B	SA003*	SRU 2	13.3	114.5	12.5	363.8	0.8	6	16.5	19.4	42.2	1.3	1	22	69	122	43.3	15.4	29.9	4.8	24
Sruhanavarnis	MAP B	SA062	SRU 2	7.6	129.2	9.5	372.5		9.7	11.2	20.9	42.4	6.3	11.8	20.8	62	132.8	53.6	21.2	30.3	7.6	27.6
Sruhanavarnis	MAP B	SA003	SRU 2	10.6	132.4	8.8	344.8		7.2	14.9	17.1	48.5	1.6	10.4	25	72.2	140.4	45.6	18.3	27.9	5.7	28.9
Sruhanavarnis	MAP B	SA086	SRU 2	10.8	156.9	9.1	357.4		5.5	13.6	22.3	54.2	3.8	10.6	16.1	73.3	138.2	20.8	13.8	20.4	6.3	31.6
Sruhanavarnis	MAP B	SA165	SRU 2	12.3	156.9	9.6	318.4		1.8	23.6	22.8	51.1	9.3	11.8	14	129.4	604.1	16.9	6.9	10.7	4.8	25.9
Sruhanavarnis	MAP B	SA174	SRU 2	12.8	133.5	10.1	307.8		1	17.6	21.6	57.8	8.6	9.5	23.4	116.1	344.2	20.9	11.1	9.6	5.8	21.9

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Nb	Zr	Y	Sr	U	Th	Pb	Ga	Zn	Cu	Ni	Cr	Rb	Ba	Ce	Nd	La	Sc	V
Sruhanavarnis	MAP B	SA106	SRU 2	10.2	139.7	12.3	331.8		7.4	19.6	20.4	56.9	7.6	8.5	18.6	111.5	563.5	71.2	24.9	26	6	29.1
Sruhanavarnis	MAP B	SA131	SRU 2	13.2	151.8	15.4	288.5		11.4	18.8	21.6	66.6	4	11.9	10.5	121.7	320.7	74.1	35.9	48.9	4.2	26.4
Sruhanavarnis	MAP B	SA019	SRU 2	16.9	140.7	13.6	377.2		9.3	16.9	23.3	62.6	2.9	10.6	18.8	94	110.1	41.7	16.3	20.7	9.6	37.5
Sruhanavarnis	MAP B	SA071	SRU 2	16.6	137.4	15.5	406.1		7.4	16	22.1	61.9	6	11.3	12.3	105.4	316.7	37.6	15.4	13.6	7.3	38.6
Sruhanavarnis	MAP B	SA117	SRU 2	14	151.7	14.5	306.8		6.4	21.3	23.4	64	3.8	10.7	14.2	129	310.8	25.5	14.9	18.8	5.3	32.1
Sruhanavarnis	MAP B	SA153	SRU 2	12.9	161.1	15.6	377.1		6.1	18.7	23	69.1	3.5	10.3	15.4	106.4	309.8	43.2	20.8	25.1	6.5	35
Sruhanavarnis	MAP B	SA008	SRU 2	15.9	161.5	13.9	388.4		4.9	14.8	20.3	60.6	8.7	9.7	19	102.1	179.7	35.8	12.6	6.1	6.4	33.1
Sruhanavarnis	MAP B	SA179	SRU 2	15.6	147.6	15.6	334.7		3.9	20	25.7	69.2	6.1	11.3	38.5	132.1	368.7	44.2	21.1	31.6	7.2	36.4
Sruhanavarnis	MAP B	SA177*	SRU 2	17.5	160.5	16.3	346.9	2.1	3.4	17.3	23.8	68.4	4.4	1.4	19	102.8	111.5	33.3	11.3	16.2	6.5	30.5
Sruhanavarnis	MAP B	SA008	SRU 2	15.2	164.2	14.1	387.7		7.2	13.3	22.9	62.2	7.1	8.3	18.9	102.4	178.5	28.3	14.8	15.3	6.5	36.3
Sruhanavarnis	MAP B	SA183	SRU 2	13.2	162.3	14	391.3		4	17.9	26.3	72.1	3	9.8	18.8	106.5	196.8	47.1	24.5	31.7	7.5	34.6
Sruhanavarnis	MAP B	SA177	SRU 2	15.4	175.3	14	331.4		3.2	17.8	28.2	77.3	4.1	9.3	16.1	106.8	108.7	37.1	16.7	12.7	9.5	34.6
Sruhanavarnis	MAP B	SA140	SRU 2	14.6	172.9	13.7	318.2		5.6	19	22.1	77.2	2.7	11.4	17.3	109.2	180	29.1	18.4	26.2	8.4	36.9
Sruhanavarnis	MAP B	SA042	SRU 2	12.4	153.8	12.4	454.9		25.5	13	24	70.9	4	12.4	19	102.4	237.8	130.7	53.6	78.3	5.8	45
Sruhanavarnis	MAP B	SA063	SRU 2	14.1	145.5	9.2	407.4		14.2	14.1	25.8	73.9	4.7	10.1	32.7	105.3	205.1	73.1	32.5	47.6	8.5	40.1
Sruhanavarnis	MAP B	SA065	SRU 2	15.7	161.3	10	415.9		11	16.2	23.3	69.1	4.1	8.5	30.5	108.5	250.5	58.9	29.2	36.5	6.8	45.9
Sruhanavarnis	MAP B	SA115	SRU 3 : PPG	8	124.3	13.9	301		10.2	26.8	16.9	29.5	4.5	11.6	8.5	158.2	989.6	62.2	35	39.5	3.8	14.7
Sruhanavarnis	SUB-MAP10	SA247	SRU 3: PPBG	8.6	118.9	11.2	409.9	2.3	8.8	26.8	13.6	36	4.4	4.1	10.8	135.2	1561	56.6	20.9	34.8	4	10.5
Sruhanavarnis	MAP B	SA095	SRU 3: PPBG	8.1	125.7	7.8	444.5		1.5	24.8	18.7	27.3	13	9.8	15.1	154.7	2050	20.5	10.1	8.2	3.8	17.3
Sruhanavarnis	MAP B	SA095	SRU 3: PPBG	8.1	121.7	7.5	444.4		2.7	25.2	18.3	30.3	13.2	9.2	10.1	154.9	2069	27.1	13.3	5	3.5	20.4
Sruhanavarnis	MAP B	SA160	SRU 3: PPBG	10.1	139.9	10.3	378.6		3.6	28.4	20.3	38.9	9.8	11.4	6.6	159.5	1832	80.9	29.9	40.4	3.3	23.3
Sruhanavarnis	MAP B	SA121	SRU 3: PPBG	10.2	142	13.2	322.7		9.6	28.2	20	40.8	6.3	11.1	13.8	161.2	1303	66.1	23.3	39.1	3.7	15.9
Sruhanavarnis	MAP B	SA158	SRU 3: PPBG	11.8	150.6	13.3	289.7		8.8	23.5	19.4	47	13.7	9.3	8.3	136.1	760.3	37.8	18.3	21.8	5.1	21.6
Sruhanavarnis	MAP B	SA211	SRU 3: PPBG	9.9	149.1	11.3	429.9		8.6	27.2	17.4	43	11.8	9.5	9.7	152.8	1913	85.2	32	42.8	4	28.1
Sruhanavarnis	MAP B	SA175	SRU 3: PPBG	9.1	150.5	10.1	391.5		7.3	28.7	20.7	45	13.6	10.9	19.1	145.7	1766	60.1	27.1	37.5	5.3	27.1
Sruhanavarnis	MAP B	SA126	SRU 3: PPBG	9.6	167	12.4	359.7		9.1	26	18.5	45	8.6	10.9	9.6	160.1	1531	79.9	31.7	34.3	7.4	24.9
Sruhanavarnis	MAP B	SA193	SRU 3: PPBG	9	148.8	11.3	418.5		8.6	25.5	22.4	37.4	7.8	9.7	11	149.3	1482	59.9	24.7	28.7	4.7	26
Sruhanavarnis	MAP B	SA182	SRU 3: PPBG	10.9	175.9	12	403.3		3.2	27.6	21	50	6.8	12.1	12.4	155.8	1626	49.8	23.9	33.6	5.7	23.9
Sruhanavarnis	MAP B	SA150	SRU 3:PSBGp	11.1	164.6	13.2	372.9		7.4	27.7	19.7	47.8	7.4	11	11.4	152.3	1468	48.1	25.3	35.5	3.2	26.2
Sruhanavarnis	MAP B	SA107	SRU 3:PSBGp	8.3	120.4	12.4	312.5		11.4	24.2	20	45.7	5.6	11.1	29.9	136.7	858.6	59.2	29.9	35	4.6	25.5
Sruhanavarnis	MAP B	SA024*	SRU 3: PSBG	8.9	81.8	10	355.1	3	4.3	22.5	15.5	25.1	4.3	2.9	5.6	117	979.7	29.8	8.6	13.4	1.6	6.5
Sruhanavarnis	MAP B	SA024	SRU 3: PSBG	8.9	100.5	6.3	355.4		6.5	21.9	15.5	23.9	6.7	12.3	12.9	119	1046	34.6	15.4	19.1	3.4	12.2
Sruhanavarnis	MAP B	SA076	SRU 3: PSBG	9.3	114.7	6.4	336.3		10.3	17.9	15.9	34.3	7.3	12.1	5.7	89.4	488.5	44.6	22.4	28.4	8.2	20.2
Sruhanavarnis	MAP B	SA038	SRU 3: PSBG	10.1	120	7.4	390.3		6.2	23	16.4	39.1	12.4	10.9	8.5	137.7	1735	55.5	25.9	34.7	6.2	22
Sruhanavarnis	MAP B	SA073	SRU 3: PSBG	10.3	112.5	9.7	355		12.8	15.5	16.1	40.3	3.3	11.1	6.1	78.9	426.7	59.7	29.4	35	3.8	19.1

LOCATION	GRID REF.	SAMPLE	GRANITE TYPE	Nb	Zr	Y	Sr	U	Th	Pb	Ga	Zn	Cu	Ni	Cr	Rb	Ba	Ce	Nd	La	Sc	V
Sruhanavarnis	MAP B	SA038	SRU 3: PSBG	10.5	120.3	7.7	389.4		7.8	22	17.5	36.2	9.8	11.2	11.4	138.3	1790	56.2	23.4	28.2	4.5	20.8
Sruhanavarnis	MAP B	SA031	SRU 3: PSBG	11.1	138.7	8.5	456.3		7.9	24.5	15.3	35.9	9.1	10.6	8.6	122.4	1789	50.4	21.6	30.8	3.9	20.9
Sruhanavarnis	MAP B	SA088	SRU 3: PSBG	9.9	137	10.1	422.1		5.7	20.2	17.8	36.7	8.5	11.1	10.4	143.2	1759	52.4	17.7	24.9	3.1	25.8
Sruhanavarnis	MAP B	SA088	SRU 3: PSBG	10.4	138.9	9.5	420.7		4.6	21.8	19.2	41.1	9.6	10.8	11.6	143.6	1775	49.3	20.7	25.6	4.9	24.1
Sruhanavarnis	MAP B	SA005*	SRU 3: PSBG	10.2	137.5	15.8	441.5	4.8	4.6	25.2	19.8	41.2	6.4	3.7	21.7	144.8	1683	46	14.8	23.2	4.2	21.6
Sruhanavarnis	MAP B	SA096	SRU 3: PSBG	12.4	165.9	9.1	471.2		6.3	20.2	19.4	45.7	8.3	10.2	21.8	138.9	1750	29.3	16.6	28.2	4.9	28.7
Sruhanavarnis	MAP B	SA005	SRU 3: PSBG	11.5	153.1	11.5	453.5		5.6	22.2	20.8	41.6	11.7	10.2	18.7	144.3	1770	44.8	17.6	22.3	3.8	26.4
Sruhanavarnis	MAP B	SA025	SRU 3: PSBG	13.1	128	12.6	408.1		3.8	23.5	20.4	49.3	5	11.3	26.7	140.1	1219	32.8	17	24.8	5	23
Sruhanavarnis	MAP B	SA025	SRU 3: PSBG	13.1	129.5	12	404.1		4.9	21.3	20.8	43.6	6.8	10.8	16.7	142.1	1194	34.8	17.3	21.6	3.6	23.5
Sruhanavarnis	MAP B	SA080	SRU 3: PSBG	13	140.7	11.2	420.2		15.6	23.2	22.5	52.6	6.9	11.7	10	137.2	1265	76.9	40.2	42.9	4.8	30.4
Sruhanavarnis	MAP B	SA014	SRU 3:VCPG	10.3	101.7	7.3	360.9		4.7	24.7	14.5	32.8	6.6	11.2	8.2	141.7	1208	29.3	13.3	18.1	4.6	19.1
Sruhanavarnis	MAP B	SA014	SRU 3:VCPG	10.4	102.8	7.4	361.3		7.1	27.5	17.4	32.3	7.7	9.6	6.1	142.6	1213	27.6	16.5	18	4.9	20.1
Sruhanavarnis	MAP B	SA016	SRU 3:VCPG	10.8	87	19.4	334.4		3.7	24.4	21.3	27.2	4.1	9.7	30.4	141.5	1029	31.6	12.3	19.1	3.4	13.9
Sruhanavarnis	MAP B	SA021	SRU 3:VCPG	11.7	113.8	10.9	474.8		0.1	28.7	16.2	33.8	12	9.8	7	159.8	2108	11.8	6.3	4.2	4.2	20.8
Sruhanavarnis	MAP B	SA083	SRU 3:VCPG	12.4	121.2	9.8	373.3		2	22.2	17.3	37.5	9.8	9	16.9	144.8	1402	11.4	5.8	6.3	3.3	28
Sruhanavarnis	SUB-MAP9	SA233	SRU 3:VCPG	11.3	155.7	11.6	346.6		11.9	26	16.9	36.3	12.6	11.7	6.5	147	1359	69.5	29.1	31.6	3.8	18.5
Sruhanavarnis	MAP B	SA222	SRU 3:VCPG	9.6	111.2	12.8	314.5		5.8	28.5	14.2	28.6	9.3	8.6	18.9	130	1115	30	14.6	10.5	4.8	16.9
Sruhanavarnis	MAP B	SA237	SRU 3:VCPG	10.8	126.1	9.6	354.9		4.5	25.2	18	29.2	11.7	12.1	17.8	132.1	1413	25.3	6.6	10.1	2.1	18.5
Trawenagh Bay	B 819023	TB2	"GWH 1"	9.4	141.8	9.2	473.3	1.9	36.6	12.9	20.6	57.2	4.3	0.3	8.4	80.8	852.9	63.5	24.3	35.3	5.1	42.6
Trawenagh Bay	B 788027	TB5	G I "GWH 3"	9.6	126.9	10.8	375.4	3.7	38.3	22.5	21	38.9	5.3	1.2	7.9	112.9	1080	32.3	13.3	22.2	3.7	21.2
Trawenagh Bay	B 7879788	TB6	G II	10	126	11.3	375.7	4.3	36.1	19.7	18.6	42.6	3.2	3.6	11.4	113.6	1066	33	14.6	20.5	1.9	20.6
Trawenagh Bay (Meen)	B 834095	TB14	G I	7.6	62.1	8.9	272.6	4.8	33.2	20.1	19.5	27.6	2.8	3.2	6.8	129.2	1554	16.1	4.5	10.1	3.9	6.4
Trawenagh Bay	B 907014	TB3	G I "GWH3"	8.9	109.3	11.5	383.4	3.4	8.3	23.7	17.2	34.5	0.8	4.2	17.2	129.4	1403	36	10.8	23.8	4	12.6



Legend

- Granodiorite (resembles that of the Thorr Pluton)
- GM 1 (fine-grained, equigranular granodiorite)
- GM 2 (medium-grained, equigranular tonalite)
- GM 3 (fine to medium-grained equigranular monzogranite)
- GM 4 (medium-grained porphyritic monzogranite)
- GM 5 (coarse grained-porphyritic monzo/ syenogranite)
- Small microgranitoid dykes
- Pegmatite

Figure 4:34

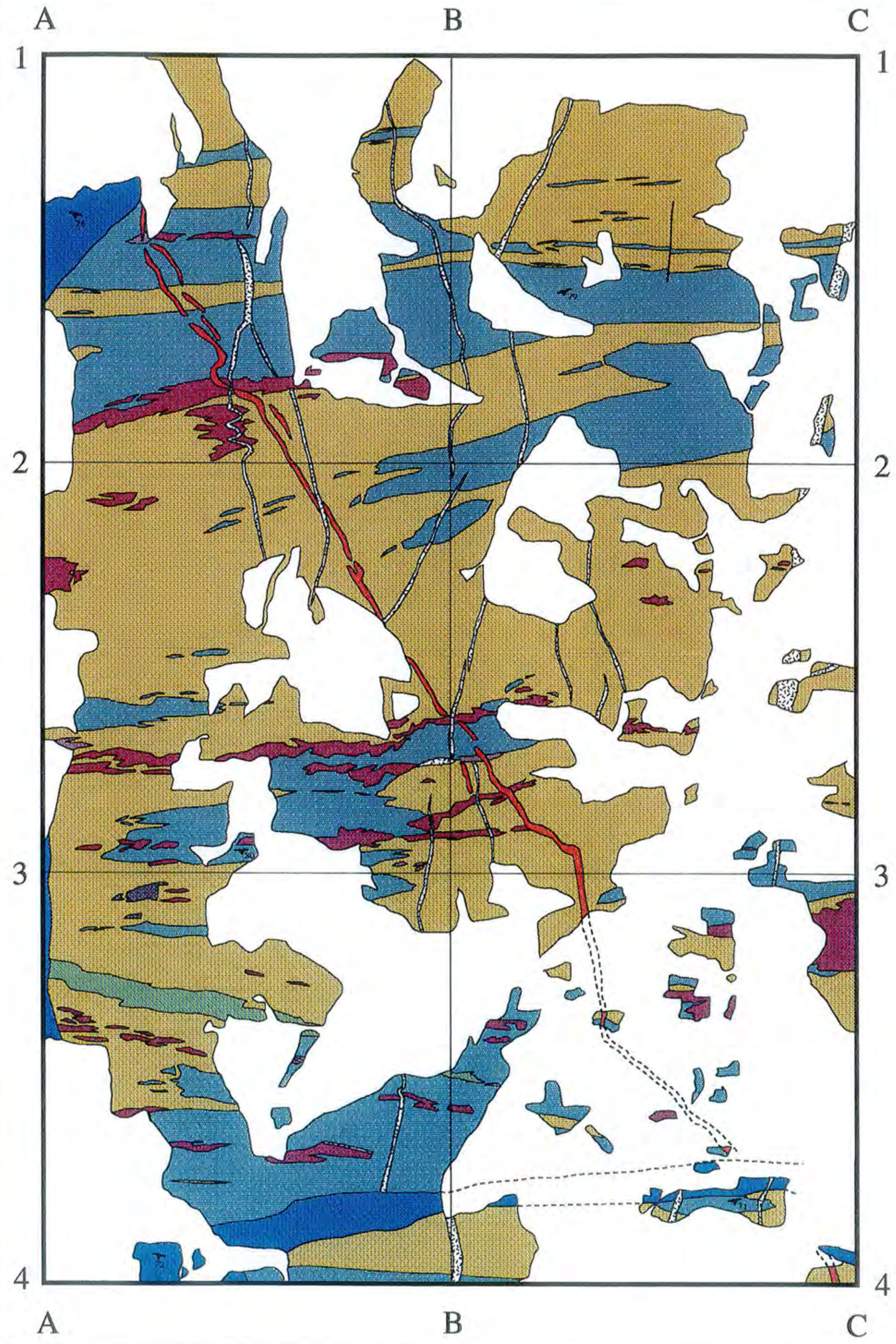
Map I

Southwest of Croaghacullin, Glendowan Mountains

0 5 10 15

Metres

N

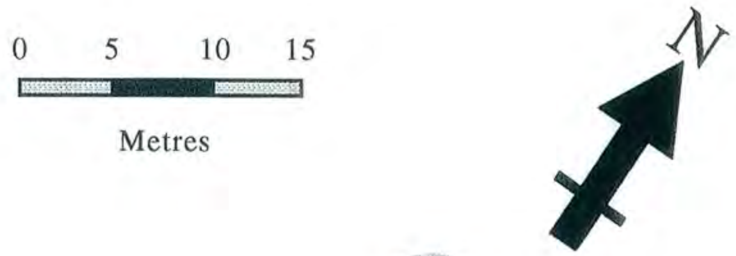


Legend

Figure 4:17

Sub-Map 9

Within the streambed of the Sruhanavarnis



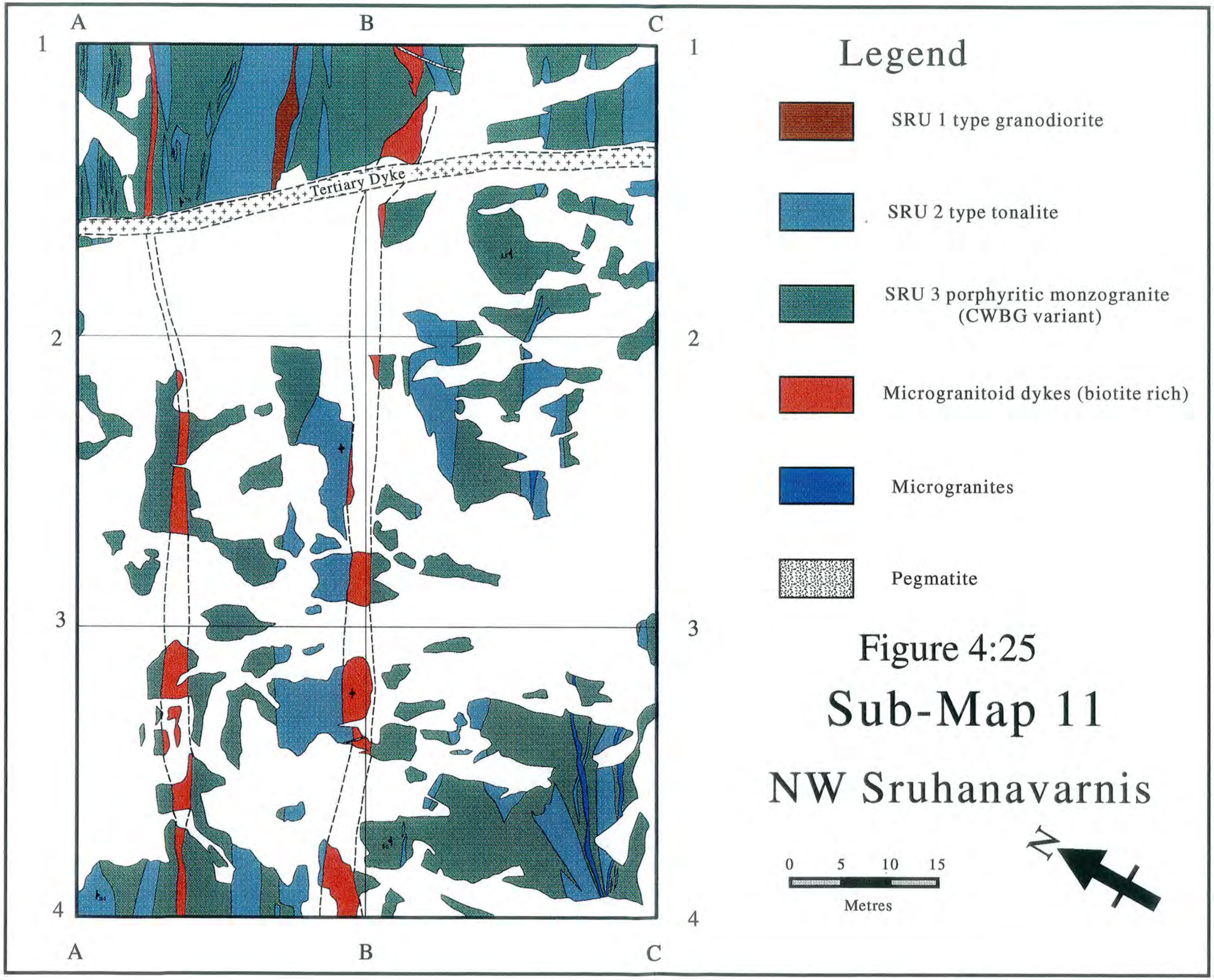
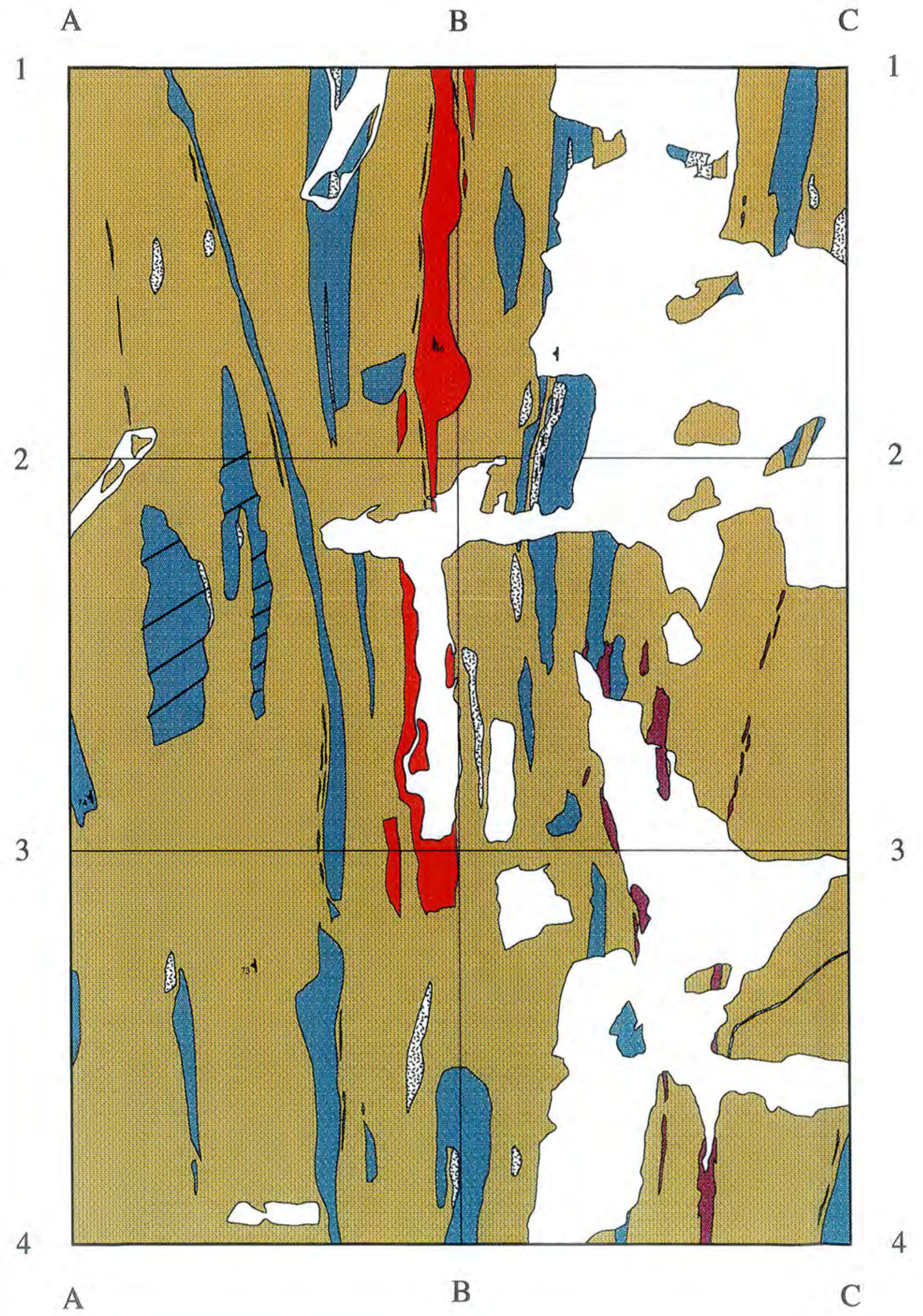


Figure 4:25

Sub-Map 11

NW Sruhanavarnis



Legend

- Pelite
- GM 2 (medium-grained equigranular tonalite)
- GM 3 (medium-grained equigranular monzogranite)
- GM 6 (Coarse-grained porphyritic monzogranite)
- Pegmatite

Figure 4:40

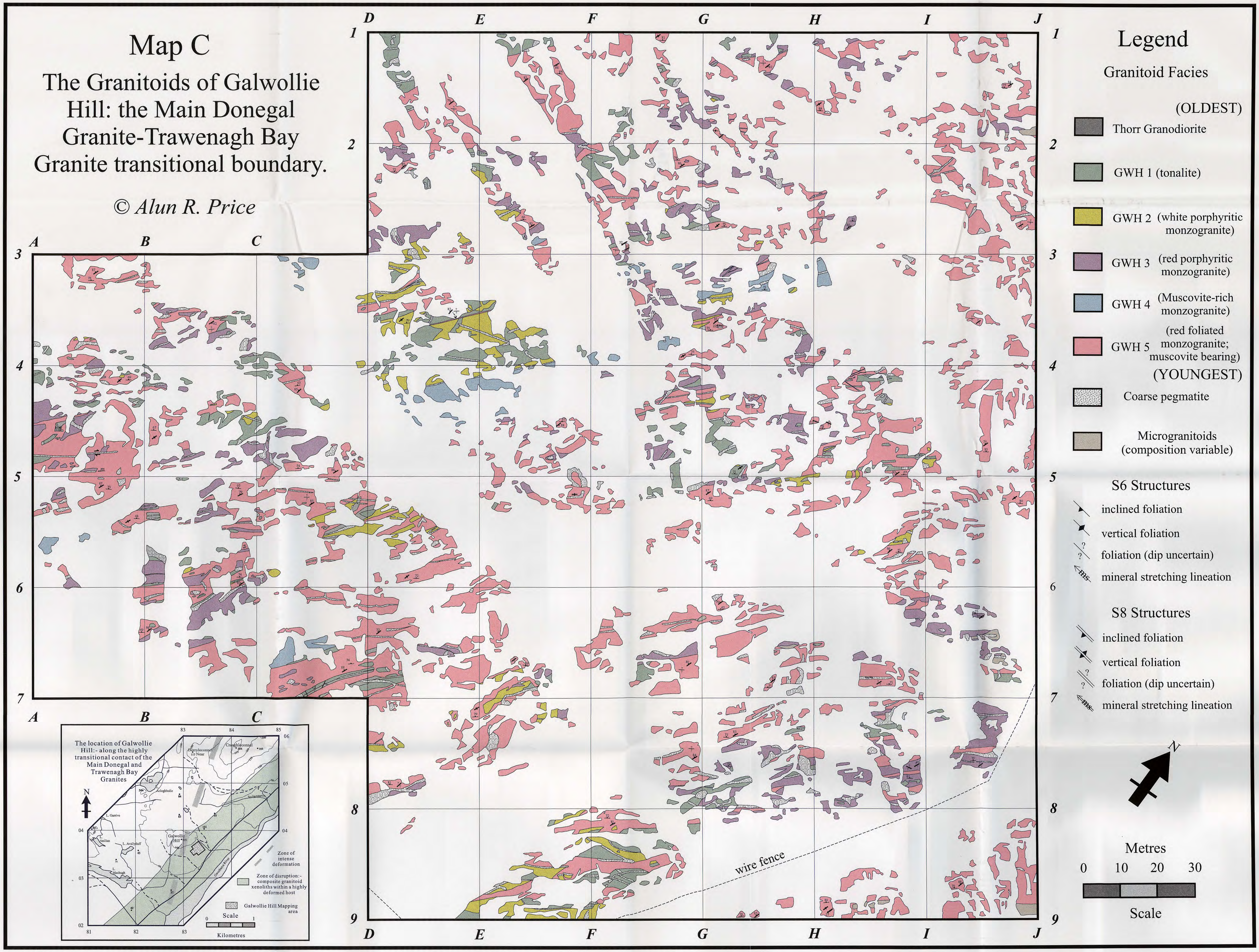
Map II

Southern flanks of Moylenanav, Glendowan mountains

0 5 10 15
Metres

N

Map D
The Internal Configuration of the Main Donegal and Trawenagh Bay plutons, NW Ireland
© 1997: Alun R. Price
(Mapping of the Trawenagh Bay pluton undertaken by W.S. Pitcher (unpublished))
THE ENVELOPE ROCKS TO THE MAIN DONEGAL GRANITE
Undifferentiated Dalradian metasediments (including metadolerites)
Thorr Pluton
Ardara Pluton
Carbane Gneiss
Meenalargan Appinite Complex
Rosses Complex
XENOLITHIC MATERIAL WITHIN THE MAIN DONEGAL GRANITE
pelite (including calc-pelite)
Quartzite
Limestone
Metadolerite
Thorr Granodiorite
Appinite
Ardara G I tonalite (and possibly some Carbane Gneiss) TXZ
THE INTERNAL GRANITOID COMPONENTS OF THE TRAWENAGH BAY GRANITE
G I:- Medium to coarse-grained, equigranular to weakly porphyritic biotite monzogranite/ granodiorite
G II:- Medium to coarse-grained, equigranular to weakly porphyritic biotite-muscovite monzogranite
G III:- Fine-grained, equigranular, biotite-muscovite monzogranite (garnet-bearing in some areas)
Muscovite-garnet aplogranites which are common towards the marginal areas of all three pulses (G I to GIII)
Coarse pegmatite in the Galwollie Hill Croaghleconnel area
534 Pts
S_6 foliation within the Main Donegal Granite
138 Pts
S_6 foliation within the Main Donegal Granite
79 Pts
S_6 stretching lineation within the Main Donegal Granite
The heterogeneous granitoid "packages" in the central regions of the pluton
"Crobane Hill-Lough Veagh package"
"Sruhanavarnis package"
"Glendowan Mountains package"
"Doocharry-Moylenanav package"
"Marginal Granitoid facies package"
The granitoids comprising the Main Donegal Pluton
The homogeneous granitoid zones
The Crockmore apophysis:- equigranular, white feldspar, biotite-rich monzogranite (fingering-out to the SW is schematically represented)
The Binaniller apophysis:- weakly porphyritic, pink monzogranites (grain-size variable and becomes progressively more variable in composition in a south-westerly direction) (fingering-out to the SW is schematically represented)
The homogeneous pink porphyritic monzogranites of the NW region of the pluton.
Legend
Boundary between different granite "packages" which is generally quite easy to observe within the field
Boundary which separates different granite "packages" (generally quite a subtle feature in the field)
Arbitary boundary based on the distribution of fine grained granodiorites
Areas of uncertainty between granitoid units
Areas of detailed mapping (1:250 scale)
Mineral alignment (dominantly S6) (inclined)
Mineral alignment (dominantly S6) (vertical)
0 1 2 3 4 5
Kilometres
Trawenagh Bay
Gweebarra Bay
Dunglow
Dunlewy
Creeslough
Glendowan
Fintown
Doocharry
Lettermacaward
Glenties



Map C
The Granitoids of Galwollie Hill: the Main Donegal Granite-Trawenagh Bay Granite transitional boundary.
© Alun R. Price
Legend
Granitoid Facies
(OLDEST)
Thorr Granodiorite
GWH 1 (tonalite)
GWH 2 (white porphyritic monzogranite)
GWH 3 (red porphyritic monzogranite)
GWH 4 (Muscovite-rich monzogranite)
GWH 5 (red foliated monzogranite; muscovite bearing)
(YOUNGEST)
Coarse pegmatite
Microgranitoids (composition variable)
S6 Structures
inclined foliation
vertical foliation
foliation (dip uncertain)
mineral stretching lineation
S8 Structures
inclined foliation
vertical foliation
foliation (dip uncertain)
mineral stretching lineation
N
Metres
0 10 20 30
Scale
wire fence
The location of Galwollie Hill:- along the highly transitional contact of the Main Donegal and Trawenagh Bay Granites
Galwollie Hill
Zone of intense deformation
Zone of disruption:- composite granitoid xenoliths within a highly deformed host
Galwollie Hill Mapping area
Scale
Kilometres


Map A
The Granitoids on Crobane Hill, Main Donegal Granite
© Alun R Price (1997)
Legend
Granitoid Facies
Thorr Granodiorite (OLDEST)
CBH1 (Medium-grained granodiorite)
CBH2 (Porphyritic granite/granodiorite)
CBH3 (Medium-grained monzogranite)
CBH4 (Fine-grained tonalite)
CBH5 (Fine-grained monzogranite)
Pegmatite (+ aplite) (YOUNGEST)
Country-rock Rafts
Quartzite
Pelite
Metadolerite
S6 Foliation
S6 Foliation (vertical)
N
Insert showing the location of Crobane Hill and its relationship to major raft-zones
L. Agannive
Slieve Snaght
L. Slieve Snaght
Addernybeg
Crockfadda
L. Nasharragh
Bingorms
R. Barra
Crocknasharragh
R254
Crobane Hill
Lough Barra
Derryveagh Raft-zone (2)
Crobane Hill Mapping area
Glenveagh Raft-zone (3A)
0 Kilometres 1
Scale
0 15 30 45 60 75
(metres)

A B C D E F G H I J K L M N O P Q R
1 2 3 4
The location of the Sruhanavarnis Valley and its relationship to the major raft-zones
The Poisoned Glen
Crockfadda
Moylenanav
Derryveagh Raft-zone, 2
Glenveagh Raft-zone, 3A
Sruhanavarnis mapping areas
Kilometres
(Contours in metres)
Sub-map Numbers (refer to text)
8 7 6 5 4 3 2 1
Legend
Dalradian meta-sediments and metadolerite "rafts"
Pelite
Quartzite
Calc-pelite
Metadolerite
Granitoid facies
SRU 1 (fine-grained granodiorite)
SRU 2 (medium-grained tonalite)
SRU 3 porphyritic monzogranites
SRU 3 (PSBG variant)
SRU 3 (VCPG variant)
SRU 3 (PSBGp variant)
SRU 3 (PPBG variant)
SRU 3 (PPG variant)
Late microgranitoid dykes
Pegmatite
S6 Foliation
S6 Foliation (vertical)
Metres
0 10 20 30
N
Map B
The Granitoids of the Sruhanavarnis Valley, Main Donegal Granite
© Alun R. Price (1997)