The structure and kinematics of the ox mountains, western Ireland: a mid-crustal transcurrent shear-zone

Jones, Christopher S.

Jones, Christopher S. (1989) *The structure and kinematics of the ox mountains, western Ireland: a mid-crustal transcurrent shear-zone*, Durham theses, Durham University. Available at Durham E-Theses Online: http://etheses.dur.ac.uk/6544/

THE STRUCTURE AND KINEMATICS OF THE OX MOUNTAINS, WESTERN IRELAND; A MID-CRUSTAL TRANSCURRENT SHEAR-ZONE.

by

Christopher S. Jones

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

October 1989.

COPYRIGHT

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work described in this thesis is entirely that of the author, except where reference is made to previous published or unpublished work.

ABSTRACT

The stratigraphy, structure and kinematics of the central and southwest Ox Mountains, a major mid-crustal shear zone, are described. The tectono-stratigraphic sequence has been established using a combination of structural and stratigraphic techniques and lithostratigraphic correlation with southern Donegal (the closest area of autochonous Dalradian). These data suggest that the Tawneyshane Formation is the equivalent of the Port Askaig Tillite and that the metasediments of the southwest and central Ox Mountains Succession represent part of the Argyll Group of the Middle Dalradian. The coarser grained rocks of the Cloonygowan Formation are tentatively correlated with the turbiditic sediments of the Southern Highland Group on the basis of lithostratigraphic similarity.

Structural analysis indicates that the Ox Mountains Succession has experienced a similar structural history to the Cloonygowan Formation although at a deeper structural level. Four distinct kinematic episodes are represented. Initial fold and fabric development (D1-D2, pre 478 Ma) was followed by sinistral transcurrent deformation, synkinematic intrusion of the Ox Mountains Granodiorite (478±12 Ma) and development of a braided system of high strain zones and tectonic slides (D3, Arenig-Llanvirn). The tectonic contact with the granulite facies metasediments of the northeast Ox Mountains is identified as a D3 tectonic slide, which dips gently to the south. This basement-cover interface strongly influences the structural geometry of the central and southwest Ox Mountains. These data suggest that at middle-lower crustal levels strike slip fault zones may be expressed as a series of high strain zones that converge both laterally and vertically. The geometry and kinematic history of the Ox Mountains is consistent with the interpretation of the inlier as the root of a major mid-crustal transpression zone. Following the cessation of transcurrent activity uplift began throughout the inlier, which is reflected by decreasing metamorphic grade and the development of conjugate folds (D4). This was succeeded by renewed sinistral transcurrent deformation of the Lough Easky and Lough Talt Adamellites (401±33 Ma), (D5, Early Devonian).

Evidence is presented that the Ox Mountains form part of the northwestern side wall of the Highland Boundary Fault Zone. Deeper levels of this structure are exposed in Ireland than in Scotland. Structural data suggest that in Ireland this structure is expressed as a major mid-crustal transpression zone that developed in response to large sinistral displacements during Arenig-Llanvirn times, followed by smaller sinistral displacements during the Early Devonian. This protracted history of sinistral transcurrent deformation can be related to terrane accretion events along the Highland Boundary Fault Zone and provides information on its early kinematic history not available in the remainder of the British Isles.

ACKNOWLEDGMENTS

I am most grateful to my supervisor Donny Hutton, for originally proposing the research topic, and for his continual interest and support during this study. His constructive criticism of the first draft of this work greatly improved the final text and provided the occupants of the coffee room with much amusement.

I wish to thank Ken McCaffrey with whom I shared two memorable field seasons in the west of Ireland, ideas on the structure of the Ox Mountains and countless pints of Guinness. Stimulating discussions with fellow postgrads, principally Richard England and Steve Jolley on subjects ranging from the genesis of granitoid magmas, strike–slip duplexes and the meaning of life are also gratefully acknowledged. Other postgrads in the Geology Department, Ian Alsop, Stuart Lake, Kevin Brown, Colin Bradshaw, Angela Hardwick Simon Hook, Jon Henton, Gerald Roberts, Dave Hunt Alick Leslie and Chris Bedford all contributed to the general air of insanity in the department and made it an extremely enjoyable place in which both to work and to play.

I am indebted to all of the members of the technical staff in the Geology department. Carole Blair typed numberous abstracts and manuscripts. Karren Gittens put up only a half hearted defence of the contents of the drawing office. Ron Lambert and George Randal prepared numberous thin sections. Dave Asberry provided additional technical support in more ways than it is possible, or prudent to mention.

In Ireland, I am grateful to Barry Long of the Geological Survey of Ireland, who generously provided outcrop maps of the southwest of the area and spent much time, both in the field and in various bars discussing the intricacies of the lithostratigraphy of the west of Ireland. I also wish to thank many friends in the west of Ireland who made my time there so memorable. In particular, I would like to thank the Maloney family of Foxford for their friendship and hospitality, and the members of the Moy Canoe Club who introduced me to the continuously replenishing pint of Guinness. Finally I would like to thank my parents, for their encouragement and support over the years.

CONTENTS

Chapter 3. Regional stratigraphic correlations, sedimentary environment and provenance. 42

Chapter 4. Deformation Chronology 56

Chapter 6. Regional tectonic synthesis and correlation 136

Chapter 7. The geometry and kinematics of the Highland Boundary Fault in W. Ireland: implications for structural geometries of accretionary tectonics 159

A geological map of the Ox Mountains is included at the back of this thesis which is intended for use with chapters 2 to 5.

CHAPTER 1

INTRODUCTION

The Ox Mountains of Co. Mayo and Co. Sligo, W Ireland (Fig 1.1), are situated along the Fair Head - Clew Bay Line, the likely continuation of the Highland Boundary Fault in Ireland, Max and Riddihough (1975). Their topographic expression is a NE-SW trending ridge extending from Castlebar in the SW, to Manorhamilton in the NE, which reaches a maximum height of 541m, and dominates the surrounding low lying areas of Carboniferous and Devonian rocks which form the margins of the inlier. Exposure in Ox Mountains is limited by extensive peat cover on the upper slopes and is best on the NW and SE flanks and in the large NW-SE trending valleys which transect the inlier. Geologically the area consists of two blocks. The central and southwest Ox Mountains, (Figure 1.1), which are the subject of the present study, contain a sequence of metasedimentary rocks both of Dalradian age, and of uncertain age, intruded by three plutons: the Ox Mountains Granodiorite, Lough Easky and Lough Talt Adamellites. This block is tectonically separated from the granulite facies rocks of the NE Ox Mountains, which have not been examined in detail.

1.1 HISTORY OF PREVIOUS RESEARCH

Previous studies in the central and southwest Ox Mountains have concentrated largely on the question of the age and stratigraphic identity of the metasediments. Conflicting views are held by two groups of workers as to whether the inlier, (except for the two tectonically separated areas of Ardvarney and Cloonygowan, agreed by both sets of workers to be Upper Dalradian), comprises Dalradian or mainly Pre-Caledonian rocks. Several attempts have been made to establish a

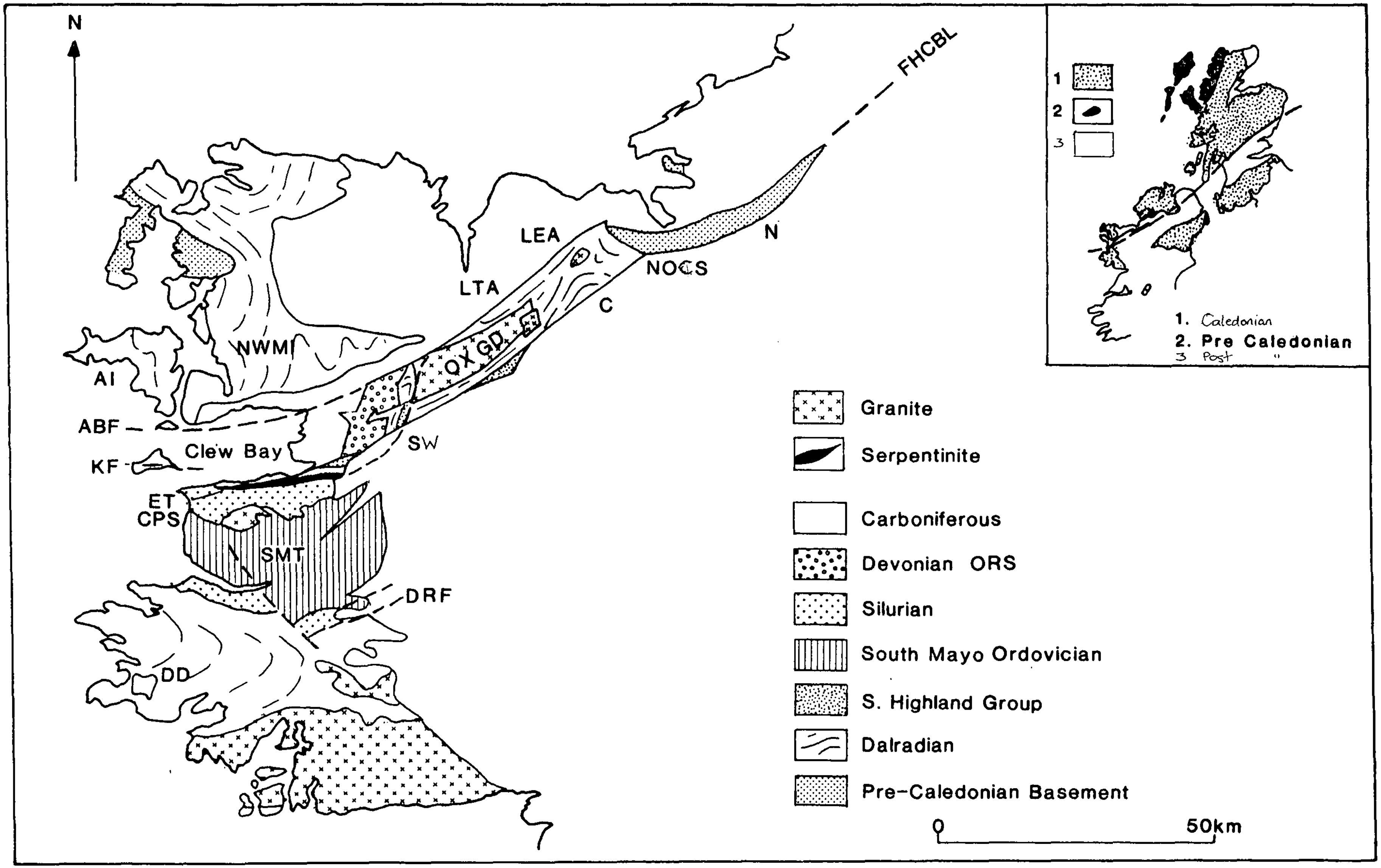

N
FHCBL
LEA
LTA
N
NOCS
C
QX GD
NWMI
AI
ABF
KF
Clew Bay
SW
ET
CPS
SMT
DRF
DD
1
2
3
1. Caledonian
2. Pre Caledonian
3 Post "
Granite
Serpentinite
Carboniferous
Devonian ORS
Silurian
South Mayo Ordovician
S. Highland Group
Dalradian
Pre-Caledonian Basement
0
50km

FIG 1.1

stratigraphic succession in both the southwest and central Ox Mountains, which have resulted in a number of different interpretations and a complex stratigraphic nomenclature summarised in Figure 2.1. The following account is not an exhaustive review of the extensive literature published on the Ox Mountains, but concentrates on the major contributions made by workers since the area was first studied in 1872, and states the current position of the main protagonists.

The earliest work was that of Symes (1872), who studied the southwestern part of the inlier and noted that a major fault, (The Ladies Brae Fault), separated the rocks of the central and southwest Ox Mountains from the rest of the higher grade rocks to the northeast. Giekie (1893) concluded that this junction represented the unconformable base of the the Dalradian overlying Lewisian, while Kinahan (1878) and McHenry (1903) both regarded the whole of the Ox Mountains as Cambro-Silurian. Dalradian affinities for the central and southwest Ox Mountains were proposed by Currall (1963), Currall and Taylor (1965), and Taylor (1969), who defined two stratigraphic successions: the northern succession consisting of all of the rocks northwest of the granodiorite, (Lough Brohly Quartzite and Attymass Group) and the southern succession south of the granodiorite, (Slieve Gamph Group and Leckee Quartzite Group) and established a stratigraphic sequence with the northern succession at the base, which they correlated with the Kilmacrennan succession of the Argyll Group in NW Donegal, (Harris and Pitcher, 1975).

The high grade rocks of the northeast Ox Mountains were subsequently examined by Lemon (1971). He considered these to be in tectonic contact with Dalradian metasediments south of the Ladies Brae. Johnstone (1975) however considered that the tectonic contact at the Ladies Brae was much less significant and correlated the whole Ox Mountains Inlier, excepting the small areas of unequivocal Upper Dalradian, with Moine.

This conclusion was supported by Phillips, Taylor and Sanders (1975) who suggested that all the rocks of the central and southwest Ox Mountains excluding those of the Cloonygowan Formation were Pre-Caledonian Moinian rocks, that are the products of amphibolite facies retrogression of the high pressure granulites exposed in the northeast Ox Mountains.

Phillips et al. (1975) considered that the entire igneous complex (Ox Mountains Granodiorite, Lough Talt and Lough Easky Adamellites) was intruded following the pre-Caledonian D2 deformation event, and cited radiometric age data for the Ox Mountains Granodiorite which suggested an emplacement age of 486 ± 6m.y., (dated by Rb-Sr, using the decay constant $^{87}Rb = 1.42{\times}10^{-11}yr^{-1}$). They regarded the intense flattening of quartz and micas in the Igneous Complex as resulting from the third deformation (D3), that was the final expression of Pre–caledonian deformation and metamorphism.

Phillips et al. stated that this was followed by the D4 and D5 events, which are correlatives of the earliest deformation observed in the Upper Dalradian rocks of the Cloonygowan Formation (Dcl, Dc2), and are therefore of Caledonian age. That is, they regard D4 and D5 as the earliest expression of Caledonian deformation which postdates the folding and metamorphism of the garnet-kyanite grade rocks of the Ox Mountains Succession, which they defined as comprising all of the rocks of the central and southwest Ox Mountains, excluding the Cloonygowan Formation. This conclusion is not accepted by the present author as the geometry and orientation of the correlated deformation events are very different. Detailed structural studies during the current research have shown that the deformation sequence in the Cloonygowan Fm. is more complex than was previously recognised, and that the Cloonygowan Fm. and Ox Mountains Succession share a common Dl-D5 deformation history, (Chapter 4, Fig 4.1).

The Ox Mountains Granodiorite that forms the core of the inlier was dated at 500 ± 18m.y. (Rb-Sr whole-rock) by Max, Long and Sonet (1976), (486 ± 12m.y. when recalculated using the decay constant $^{87}Rb = 1.42{\times}10^{-11}yr^{-1}$), who described its emplacement as syn-late kinematic. Independant dating by

Pankhurst et al. (1976) gave a similar Rb-Sr whole-rock age of 487 ± 6 m.y., (recalculated at 478 ± 12 m.y. using the new decay constant). Both groups of workers assumed that the Lough Talt and Lough Easky Adamellites were of the same age, although they did not separately analyse samples from either of these plutons.

At this time, Long and Max, working in the southwest and central Ox Mountains, while accepting the deformation sequence and correlation between the deformation chronologies of the Ox Mountains Succession and the Cloonygowan Formation proposed by Phillips et al. (1975), suggested that the entire Ox Mountains inlier south of the Ladies Brae might be Dalradian, and that it underwent Caledonian polyphase deformation and regional metamorphism similar to adjacent Dalradian inliers at around 500Ma. They considered that the variety of lithologies and particularly the presence of marbles and metavolcanic units within the central and southwest Ox Mountains to be characteristic of rocks of a Dalradian age, and defined a coherent tectono–sedimentary succession. They regarded the Attymass Group of Taylor (1969) as being equivalent to part of their Cappagh Formation and cited way up evidence from the Leckee Quartzitic Formation to support a stratigraphic succession, which included two major metavolcanic units, the Newantrim Member and the Callow Formation. The Tawneyshane Marble Member, believed to be exposed in the core of an F2 antiform within the Leckee Quartzitic Formation was regarded as lying at the base of the succesion. They then correlated this succession with the Appin and Argyll Groups of the Dalradian of Connemara, Harris and Pitcher (1975), Long (1974), Long and Max (1977).

In their discussion of Long and Max (1977), Phillips et al. (1977) rejected these conclusions and re-iterated their view that the lithological sequence was similar to Moine rocks, and that all of the amphibolites in the Ox Mountains were of intrusive origin. They acknowledged however that lenses of marble were present in the Ox Mountains, which they conceeded were rarely found in the Moinian rocks with which they correlated the Ox Mountains.

Andrews, Phillips and Molloy (1978), studied part of the central Ox Mountains between Lough Talt and the Ladies Brae and argued that the absence from the northwest side of the inlier of the coarse garnetiferous pelite that is continuously developed along the contact between the Ummoon Formation and the Leckee Quartzitic Formation, and the differences between the semi–pelite and psammite of the Lismoran Formation on the southeast side of the inlier and the pelite and semi–pelite of the upper part of the Attymass Group was inconsistent with the Attymass Group being equivalent to the Cappagh Formation of Long and Max (1977). They therefore concluded that the Attymass Group underlies the Leckee Quartzic Formation. Andrews et al. (1978) developed the deformation chronology established during Phillips' earlier work in 1975. They considered that the amphibolites which Long and Max (1977) regard as extrusive, to have been intruded at a late stage of D2, prior to the mid-amphibolite facies MP2 peak of regional metamorphism. They regarded the Ox Mountains Granodiorite, Lough Easky and Lough Talt Adamellites to be of Caledonian age and emplaced coevally with the D5 and D6 deformation that took place under greenschist facies conditions. Therefore the Caledonian-Grampian cycle is younger than the D1-D4 cycle in the Ox Mountains sequence, that comprises all of the rocks of the Central and southwest Ox Mountains, excluding the Cloonygowan Formation. Andrews et al. (1978), stated that if the D1-D4 deformation was 'early' Grampian within an unusual, perhaps littoral, facies of the Dalradian, tectonic uplift and cooling could have taken place before the rocks were juxtaposed with the Upper Dalradian rocks of the Cloonygowan Formation and reworked by the full Dalradian cycle in the younger rocks. This would however be a unique situation within the Dalradian of the British Isles and would imply that there should be a major unconformity in the Dalradian, corresponding to the uplift and cooling stage of the Ox Mountains sequence, which had not been found. They concluded that the D1-D4 cycle is pre-Caledonian and therefore favour the first two of the four possible tectono-thermal models which they present, (Figure 1.2), and are summarized below.

1. Reworking Hypothesis: The Granulite facies rocks of the northeast Ox Mountains have experienced an 'early' pre–Grampian orogenic cycle. The D1–D4 deformation in the Ox Mountains sequence is part of a 'later' pre–Grampian cycle. The Grampian tectonothermal cycle is fully represented by the D5 and D6 low greenschist facies retrogression in the central Ox Mountains sequence and the Dc1–Dc2 deformation in the Cloonygowan Formation.

2. Reworking hypothesis with diachronism: The Ox Mountains sequence of the northeast and central Ox Mountains are all part of the Moinian succession. They have experienced a single pre–Grampian orogenic cycle, with only the northeast Ox Mountains having been subjected to granulite facies conditions. The simpler structural history and lower grade of the central Ox Mountains is explained by tectonic juxtaposition during a diachronous pre–Grampian orogeny. The Grampian cycle is confined to shear zones, the Cloonygowan Formation and the Igneous Complex.

3. Model of moderate diachronism: The granulite facies deformation of the northeast Ox Mountains is pre–Grampian. The D1–D4 amphibolite facies deformation in the central Ox Mountains and the Dc1–Dc2 deformation in the Cloonygowan Formation are both Grampian. Thus an early Grampian cycle has been reworked by a late Grampian cycle as a consequence of tectonic juxtaposition in a diachronous orogeny.

4. Hypothesis of extreme diachronism: All three tectonothermal cycles are Grampian. Recycling has been produced by multiple tectonic juxtaposition during a highly diachronous Grampian orogeny.

The metamorphic studies of Yardley, Long and Max (1979), and Yardley (1980), demonstrated that the central and southwest Ox Mountains experienced

FIG 1.2

OUNTAINS

PHILL

PRESENT STUDY

1. Reworking hypothesis

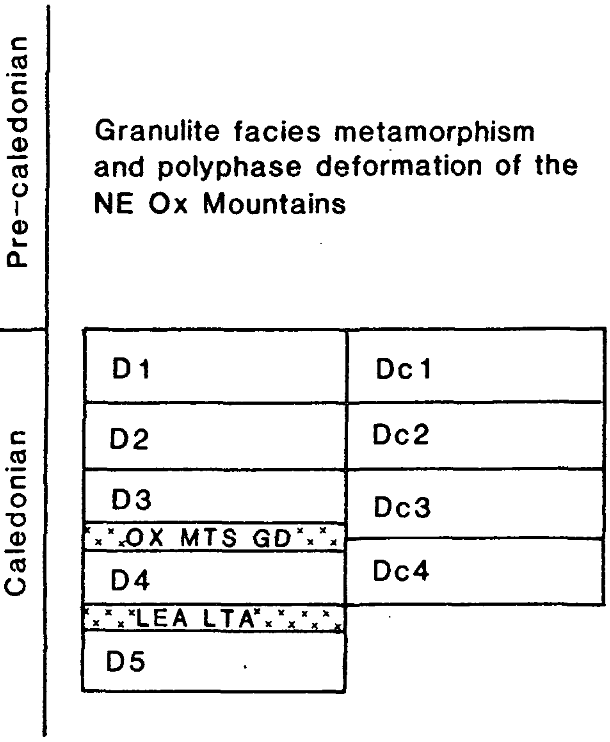
The Grampian thermo-tectoni is fully represented by the cycle Cloonygowan Fm.- Ox Mountains I Complex, and low greenschist facie gression and cataclasis in the Ox tains sequence of the central Ox tains. The earlier prograde (D1-D4 in the Ox Mountains sequence is a pre-Grampian orogenic cycle. T mulite facies cycle in the NE Ox Mo is an even earlier pre-Caledonian o cycle.

2. Reworking hypothesis with diach

The central and northeast Ox tains are part of a single stratigraph cession, probably Moine. They hav through a single pre-Grampian o cycle with only the metasediments NE Ox being *subjected* to granul cies conditions and more complex tural events. The simpler structural and amphibolite facies metamorph the central Ox Mountains, and its print on the granulite facies rocks, explained by tectonic juxtaposition a diachronous pre-Grampian orogen Grampian cycle only reached low schist facies conditions and is largel fined to shear zines, the Cloonygowa and the igneous complex in the cent Mountains.

3. Hypothesis of moderate diachroni

The amphibolite facies D1-D4 c the central Ox Mountains and the I cycle of the Cloonygowan Fm. are Grampian. An early Grampian cyc been reworked by a late Grampian c a consequence a tectonic juxtapositio diachronous orogeny. The Granulite cycle of the NE OX is pre-Grampian



The pre-caledonian granulite facies metamorphism and deformation of the NE Ox Mountains is succeeded by a single Caledonian orogenic cycle in that deforms the Cloonygowan Fm. synchronously with the amphibolite facies deformation in the Ox Mountains Succession.

The difference in metamorphic grade is considered to reflect deformation that occurred at different levels in the crust.

a single progressive Barrovian metamorphism from garnet to kyanite zone and suggested that the tectonic contact between the granulites of the northeast Ox Mountains and the Barrovian kyanite zone rocks of the central Ox Mountains is analogous to that in southern Donegal and that the conditions of metamorphism of the Dalradian staurolite-kyanite schists was the same in S. Donegal (Lough Derg) and the central Ox Mountains.

Yardley and Long (1981), recognised that the Lough Easky and Lough Talt adamellites were relatively undeformed compared to the granodiorite, and that the conspicuous thermal aureole developed around them contrasted with the lack of an obvious aureole around the granodiorite. Using five independent criteria they determined that the conditions of metamorphism in the inner aureole of the Lough Easky Pluton were 595 ± 30°, 2.5 ± 0.5 kb, which corresponds to an emplacement depth of approximately 8km and that the Lough Talt Pluton was emplaced at slightly higher pressures. They also noted that a single sample from the Lough Easky Adamellite analysed by Pankhurst et al. (1976), fell below the isochron for the other Ox Mountains granodiorite samples. They therefore suspected that both adamellites might be of a younger age than the granodiorite. This was confirmed by the Rb-Sr dating of Long, Max and O'Connor (1984) which showed that they were significantly younger at 401 ± 33Ma. Andrews (1984), studied the Lough Easky Adamellite and concluded that it was deformed under greenschist facies conditions and was intruded into a pull-apart caused by a two kilometer offset perpendicular to strike, in a NNE trending dextral shear zone, which he termed the Lough Easky Slide.

Hutton and Dewey (1986) recognised that much of the deformation in the Ox Mountains was the result of intense sinistral transcurrent motion and suggested that the broad antiformal cross-section of the Ox Mountains represented a flower structure that developed in a sinistral transcurrent shear zone operating slightly later than the intrusion of the granodiorite. They related this deformation to the docking of allocthonous terranes lying to the south, against the inlier.

The geochemical study of Winchester et al. (1987) is the most recent published work on the area. These authors found geochemical differences between two metabasalt units that occur within the metasedimentary sequence and are separated by a syn-metamorphic dislocation, (the Glennawoo slide; Taylor 1969). Winchester et al. demonstrated that the metabasalts south of the slide (the Callow Member of the Lismoran Formation) have trace element compositions similar to mid ocean ridge basalts, whereas metabasalts north of the slide, (the Newantrim Member of the Ummoon Formation) have trace element compositions similar to continental tholeiites and tholeiites of ocean island basalt association. They did not regard the great similarity of the lithological characteristics of the two seperated units (the Lismoran and Ummoon Formations) as significant and concluded that the two groups of metabasalts were erupted in different crustal settings and proposed the Glennawoo slide as a major terrane boundary. This conclusion was discussed in detail by Jones and Leat (1988), who stated that the geochemical differences between the metabasalts documented by Winchester et al. can be satisfactorily explained by their generation by different degrees of partial melting of the same asthenospheric mantle. They concluded that since the Glennawoo Slide separates rocks whose structural and stratigraphic histories are indistinguishable, there is no basis for regarding the Glennawoo Slide as a major terrane boundary.

Research currently in progress includes an attempt by W.E. Taylor (Luton Polytechnic) to see if the stratigraphic horizons have a unique geochemical signature that can be used to identify them. K. McCaffrey (Durham) is studying the emplacement and deformation of the Ox Mountains Granodiorite, the Lough Talt and Lough Easky Adamellites, and J. Reavy (Liverpool) is examining the trace element geochemistry of the igneous lithologies. Detailed stratigraphic studies by B. Long (Geological Survey of Ireland) are also continuing.

In summary, much of the previous research in the southwest and central Ox Mountains has been directed towards determining the age and stratigraphic identity of the metasediments. Currently conflicting views are held by two groups of workers, (Fig 1.2). One group, principally Phillips and Andrews, consider that the rocks of the entire inlier, excepting the Cloonygowan Formation are pre-Caledonian in age, and favour a complex tectono-thermal history involving three separate phases of sedimentation, metamorphism and deformation. The other group of workers lead by Long and Max, believe that the rocks of the central and southwest Ox Mountains are of Dalradian age and that two orogenic episodes are represented. The first episode which was of pre-Caledonian age deformed and metamorphosed the NE Ox Mountains, was followed by Caledonian metamorphism and deformation which affected the rest of the inlier.

1.2 RESEARCH OBJECTIVES.

The principal objectives of this research were :

1. To establish the age and stratigraphic identity of the Ox Mountains metasediments.
2. To develop a structural and kinematic model for the area, and specifically test the hypothesis that these rocks are part of the exhumed mid-crustal side wall to the Highland Boundary Fault, (Hutton and Dewey, 1986).
3. To develop data and ideas which are relevant to the general problem of processes and deformation geometries in mid-crustal strike slip shear zones and their relationship to granitic magmatism.

Structural, kinematic and stratigraphic data collected during 14 months fieldwork conducted during 1986-1988, when the metasediments of the inlier which in total covers approximately 440 Km2) were mapped on a scale of 1:10,560, have been used to compile 1:25,000 structural and stratigraphic summary maps. These field data provide the basis of the structural, kinematic and stratigraphic analysis that follows.

CHAPTER 2

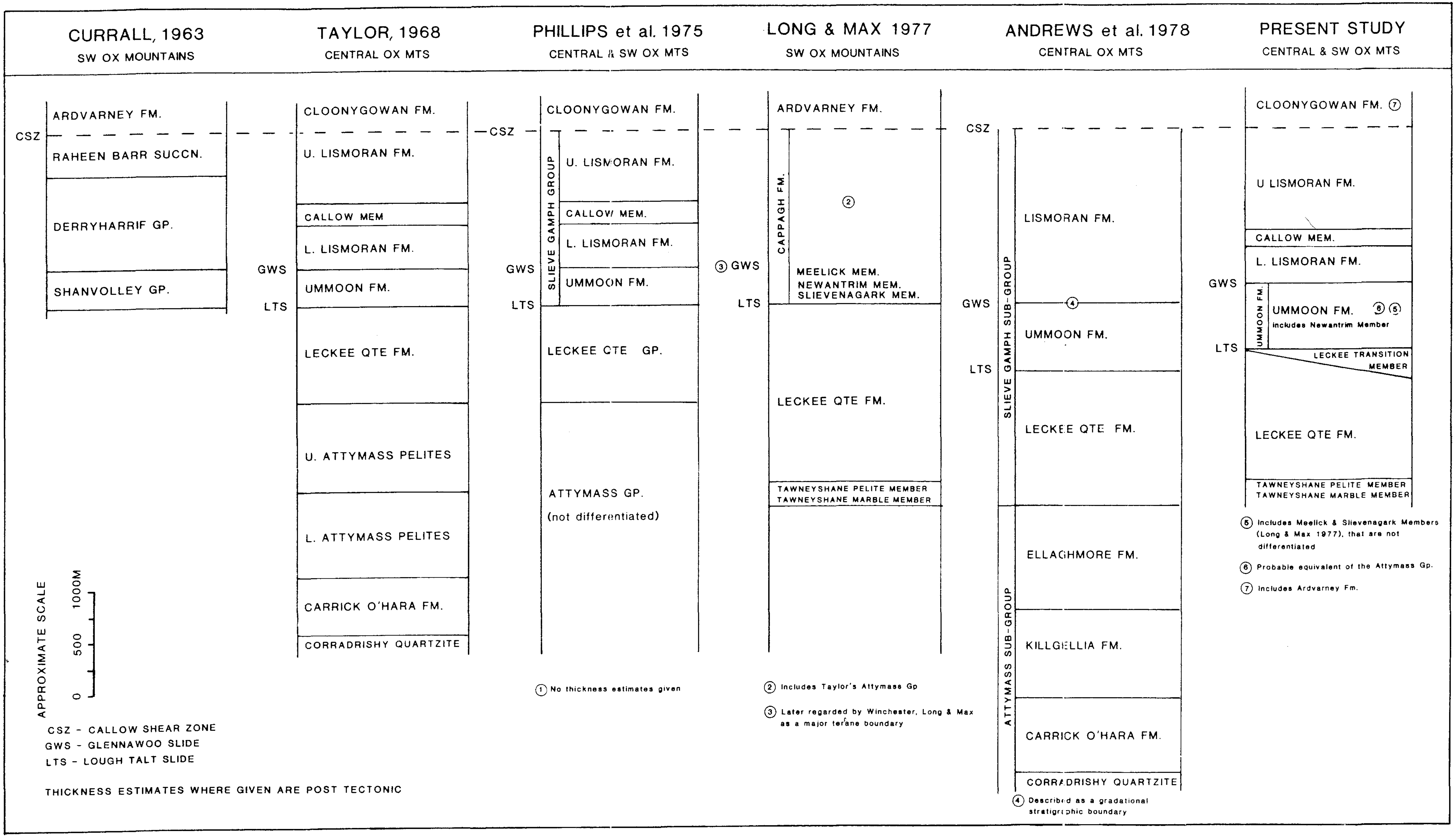
STRATIGRAPHY

Previous stratigraphic studies in separate areas of the central and southwest Ox Mountains have lead to a number of different interpretations of the succession and a proliferation of names for equivalent stratigraphic units. There is little agreement in the literature on whether a stratigraphic sequence can be established in the Ox Mountains, and if so what way up it is (fig 2.1). Two principal stratigraphic problems must therefore be addressed. The first of these is to determine if it is possible to establish a tectono–stratigraphic sequence and by determining its way up, to demonstate the original stratigraphic succession.

In many areas of the Caledonides a stratigraphic sequence can be established using criteria such as graded and cross-bedding to determine the way up of strata. In areas that have experienced a polyphase deformation history, it is often necessary to remove the effects of a commonly geometrically complex series of folds, before establishing the stratigraphic sequence. The concept of facing (Shackleton 1958), which is defined as the resolution of the way up of the beds into the axial plane of a fold, is a useful means of describing the relation between folding and the original geometry of bedding and is commonly used in such areas. The problem is more acute when the sequence is dismembered by structural breaks such as tectonic slides, or when intense deformation transposes bedding and sedimentary structures. Alsop (1987), demonstrated in Central Donegal that the Central Donegal Slide separates two stratigraphic packages which ubiquitously young away from the slide zone. This clearly illustrates that it is not possible to extrapolate the way up of a stratigraphic succession across structural breaks.

FIG 2.1

STRATIGRAPHIC NOMENCLATURE IN THE OX MOUNTAINS



CSZ - CALLOW SHEAR ZONE
GWS - GLENNAWOO SLIDE
LTS - LOUGH TALT SLIDE

THICKNESS ESTIMATES WHERE GIVEN ARE POST TECTONIC

① No thickness estimates given

② Includes Taylor's Attymass Gp

③ Later regarded by Winchester, Long & Max as a major terane boundary

④ Described as a gradational stratigraphic boundary

⑤ Includes Meelick & Slievenagark Members (Long & Max 1977), that are not differentiated

⑥ Probable equivalent of the Attymass Gp.

⑦ Includes Ardvarney Fm.

In the Ox Mountains all of these difficulties are present; a polyphase deformation history combined with intense transcurrent deformation has modified bedding such that it is often totally transposed into the pervasive D3 tectonic fabric. This has almost completely eliminated sedimentary structures which have only been observed in the Leckee Quartzitic Formation and in the low-grade rocks of the Cloonygowan Formation. Both of these units are separated from the rest of the inlier by tectonic breaks. Since it is not possible to extrapolate data across these contacts, it is necessary to use other criteria to establish a stratigraphic sequence. Three additional techniques have been used in the present study.

1. The bilateral symmetry of units in the absence of tectonic breaks may be used to infer the existence of folds. Thus the identification of the Tawneyshane Pelite Member on either side of the Tawneyshane Marble Member at Lough Anaffrin (M 1678 9621), indicates that the later is exposed in the core of a fold that predates the D3 deformation, whose associated folds consistently verge to the northwest in this area. Minor structures associated with D2 have been transposed into S3, and therefore cannot be used to confirm the existence of this fold.

2. The identification of original stratigraphic transitions between units can be used to establish stratigraphic continuity between units in a sequence in which structural breaks are commonly developed.

3. Most importantly, unique stratigraphic marker horizons such as the Port Askaig Tillite, can be used to correlate with areas such as S. Donegal where the stratigraphic sequence is more clearly defined.

The second stratigraphic problem is to determine the stratigraphic relationship between the rocks exposed northwest of the Ox Mountains Granodiorite and those exposed to its southeast. Incomplete exposure has resulted in two views. One group of workers, principally Long and Max, regard the rocks exposed northwest of the granodiorite as equivalent to those of the Cappagh formation exposed on the southeastern side of the inlier. The second group of workers principally Andrews and Phillips, regard the rocks exposed southeast of the granodiorite as stratigraphically underlying those on the northwest side of the granodiorite. Normal techniques of lithostratigraphic correlation combined with structural analysis have been used to try to solve this problem.

The inlier has been divided into six areas for ease of lithological description (Figure 2.2). Area 1 comprises the high grade metasediments exposed on the on the southeast of the inlier. Area 2 is located at Lough Talt, north of the northern end of the granodiorite, where the geographical separation between the rocks exposed on the northwestern side of the inlier and those on the southeastern margin is a minimum. This area therefore provides critical evidence for the correlation between areas 1 and 3, which comprises the metasediments exposed on the northwestern side of the granodiorite between Ilnaglashy and Tawnaneilleen. Area 4 is located west of the Knockaskibole fault, and comprises metasediments that are correlated with the rocks of area 1, exposed east of the fault. Area 5 is also located west of the Knockaskibole fault immediately south of area 4, with which it is in fault contact. The rocks of area 5 are also in tectonic contact with those of the Cloonygowan Formation of area 6b to the east. Area 6 comprises the low grade rocks of the Cloonygowan Formation that are exposed in three localities; at Cloonygowan (area 6a), southeast of area 1, at Ardvarney (area 6b) west of the Knockaskibole fault and near Westport (area 6c).

The stratigraphic sequence established in the following sections, has been developed using a combination of all of the above techniques, and is summarised in Figure 2.3. This differs significantly from that of previous workers which is summarised in figure 2.1. The stratigraphic interpretation of the southwest and central Ox Mountains is illustrated in Figure 2.4 (contained in the map pocket inside the back cover of this volume).

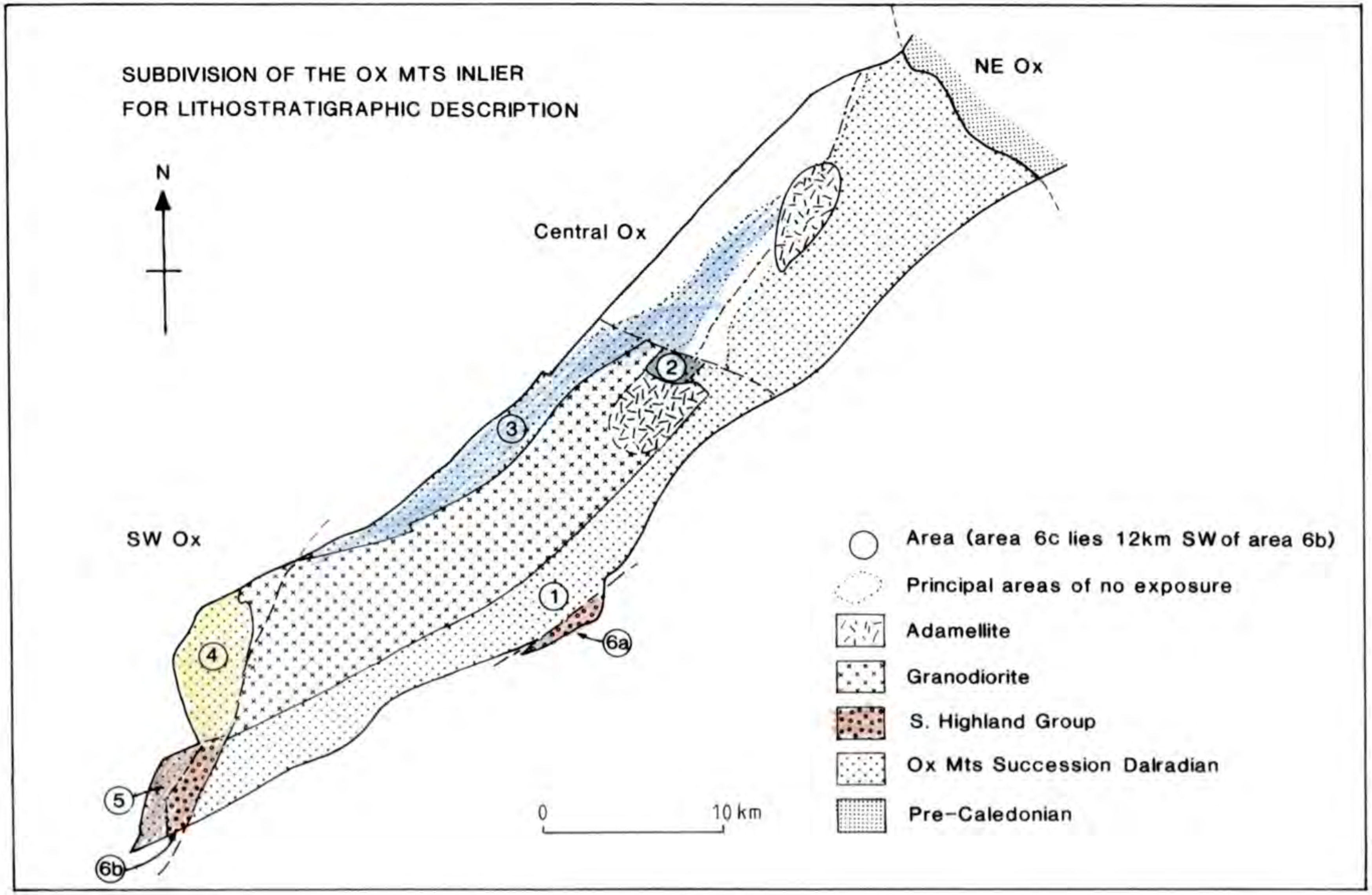
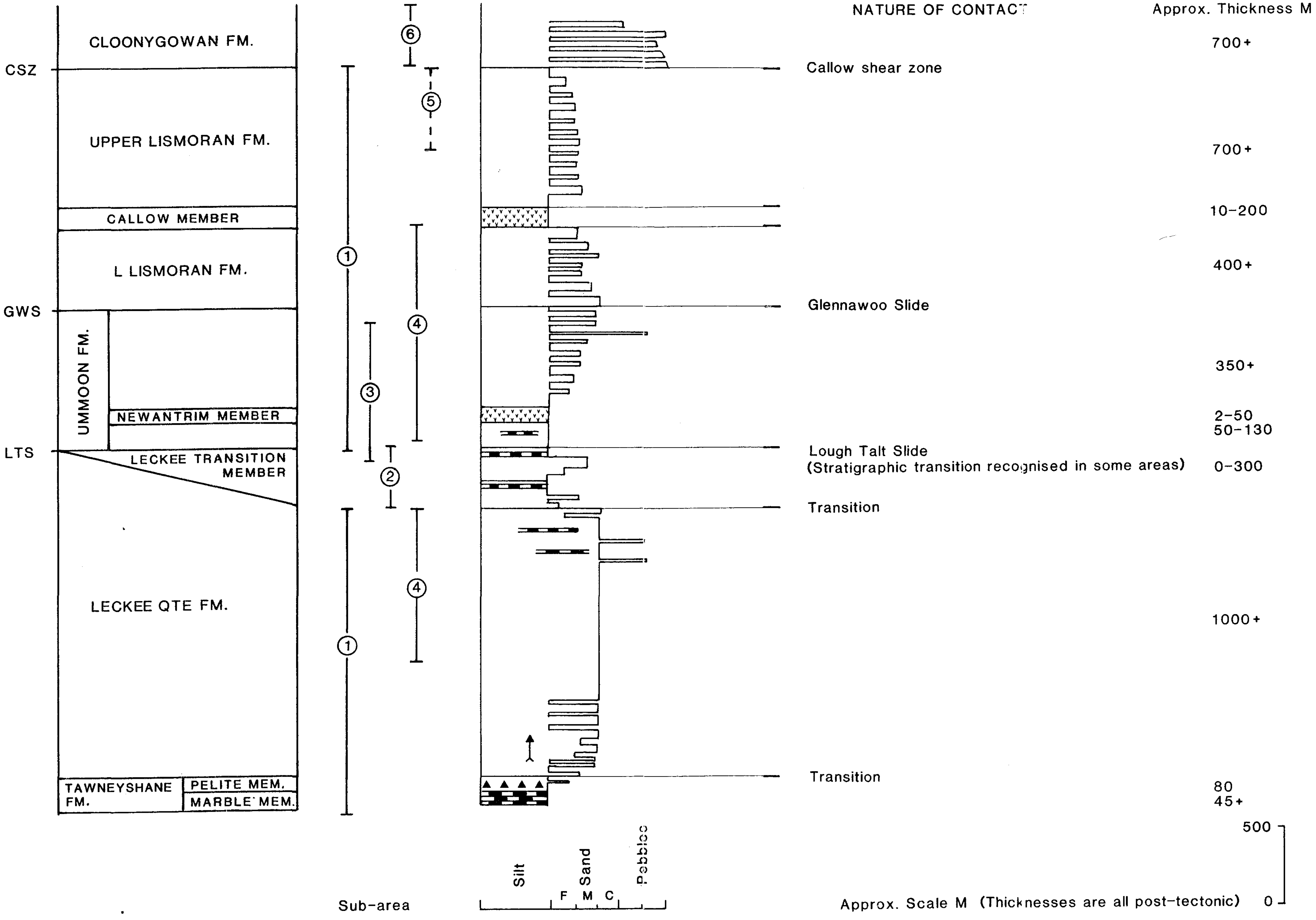

SUBDIVISION OF THE OX MTS INLIER FOR LITHOSTRATIGRAPHIC DESCRIPTION

FIG 2.2

FIG 2.3

Composite stratigraphic sequence in the central and southwest Ox Mountains



2.1 STRATIGRAPHY SE OF THE GRANODIORITE : AREA 1

Area 1 is the largest area of outcrop in the central and southwest Ox Mountains. The metasediments of the area are highly deformed and are dismembered by tectonic slides. However they provide important data which have been used to establish a tectono-stratigraphic sequence (Figure 2.3), which is correlated with area 3, using area 2 as the link between them.

2.1.1 Tawneyshane Formation

Exposure of this formation is confined to an area at Lough Anaffrin (M 1678 9621), where it is in contact with the Leckee Quartzitic Formation on its northwest and southeast sides. Two possibilities exist to explain this outcrop pattern. The first, is that the Tawneyshane Fm is exposed in the core of a fold that is of pre-F3 age, (F3 folds consistently verge to the northwest in this part of the section). This is considered by the present author to be more likely than the alternative option that the Tawneyshane Fm represents a lens within the Leckee Quartzitic Fm, as this would require the bilateral symmetry of the former unit to be accepted as coincidence. Unfortunately minor structures related to the D2 deformation have been totally transposed into the S3 fabric in the area, therefore no structural data were observed to confirm this model.

2.1.1.1 Tawneyshane Marble member: The Tawneyshane Marble Member consists of white tremolitic marble interbedded with green serpentinous marble and white to pale green or grey calc-silicate schist composed of dolomite, calcite muscovite, biotite, phlogopite, chlorite, diopside, tremolite, actinolite, sphene and tourmaline. The unit is only exposed in one small (2m x 4m) outcrop on the north shore of Lough Anaffrin (M 1678 9621), but exploratory drilling and trenching by the Geological Survey of Ireland (Long et al. 1981), allowed it to be traced for 500m along strike, (fig 2.5). The base of the member is not exposed, therefore the observed thickness of 80m is a minimum. A narrow transitional boundary in

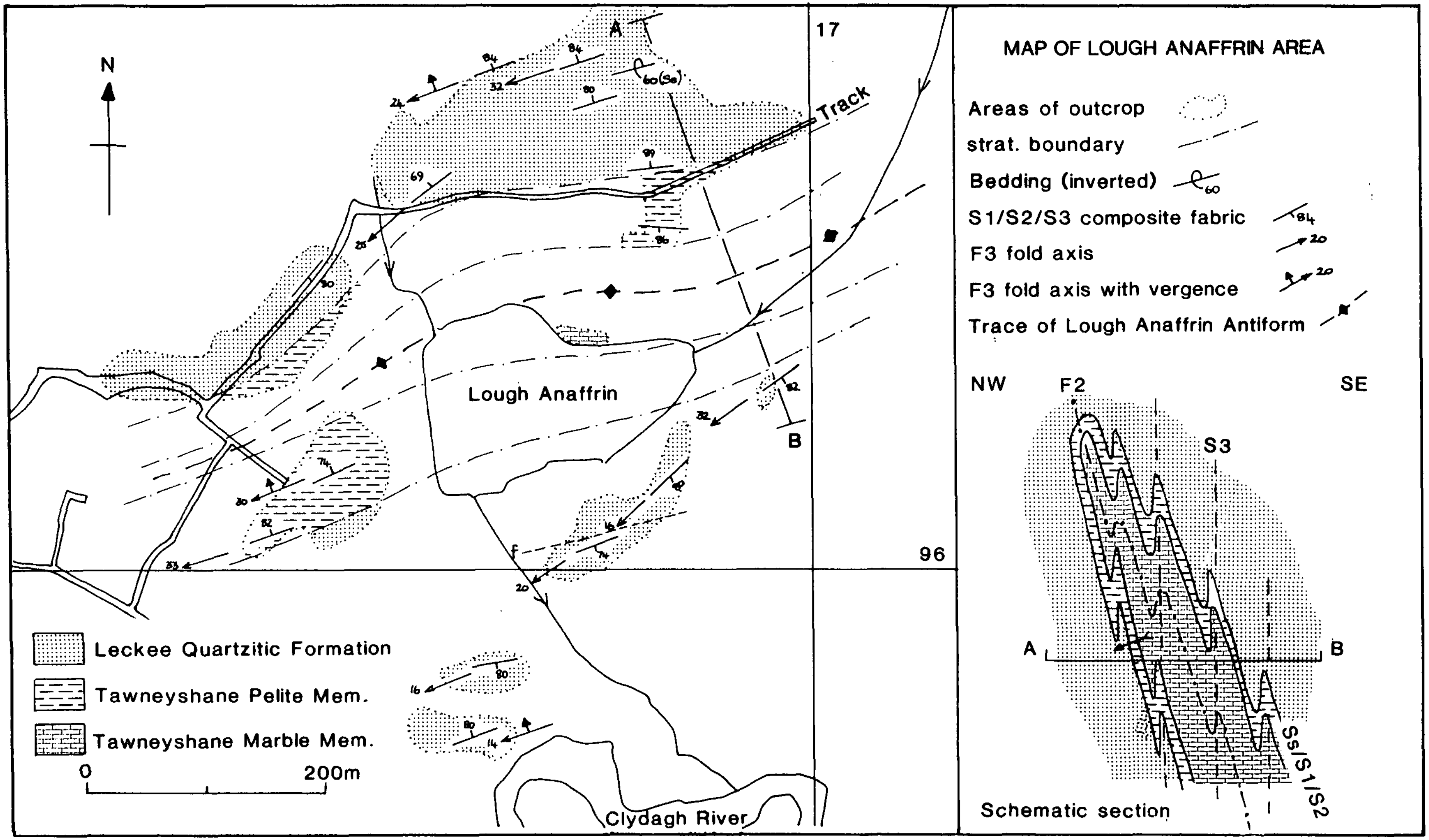

MAP OF LOUGH ANAFFRIN AREA
Areas of outcrop
strat. boundary
Bedding (inverted) 60
S1/S2/S3 composite fabric 84
F3 fold axis 20
F3 fold axis with vergence 20
Trace of Lough Anaffrin Antiform
N
17
96
A
B
Track
Lough Anaffrin
Clydagh River
f
84
84
32
24
80
60(Se)
69
99
86
25
30
74
30
82
33
82
32
80
16
74
20
80
16
80
14
Leckee Quartzitic Formation
Tawneyshane Pelite Mem.
Tawneyshane Marble Mem.
0
200m
NW
F2
SE
S3
A
B
Ss/S1/S2
Schematic section

FIG 2.5

which thin pelite interbeds are present can be recognised between this and the Tawneyshane Pelitic Member.

2.1.1.2 Tawneyshane Pelitic Member: The Tawneyshane Pelitic Member is poorly exposed on both sides of the Tawneyshane Marble Member. East of Lough Anaffrin (M 1690 9621), the marble passes into massive poorly banded greenish-weathering pelite and semi-pelite, composed of muscovite, quartz, plagioclase, biotite, chlorite and tourmaline (Plate 2.1). The proportion of semi-pelite gradually increases upwards towards a narrow band which has been found to contain a number of small (0.5cm) psammitic pebbles. This occurs 20m from the base, (Figure 2.5). With increasing thickness and proportion of psammitic and quartzitic bands, the unit passes transitionally into the Leckee Quartzitic Formation to the southeast.

The northern boundary of the Tawneyshane Pelitic Member is also exposed on a small hill 50m northwest of Lough Anaffrin, (M 1660 9616), (Figure 2.5), where thin, (10cm) psammitic bands are interbedded with approximately 15m of brownish weathering pelite.

A large number of isolated lenses of granitic material similar to those described by Long and Max (1981) from the east side of Lough Anaffrin, and also composed of quartz, feldspar, muscovite and tourmaline, have been discovered here in a band 2m wide (Plate 2.2). Cutting equipment was used in the field by the present author to section bands containing granitic material. In all of the samples examined the granitic material was revealed to be boudinaged veins, and not detrital in origin.

As will be further discussed in chapter 3, the lithology of the Tawneyshane Pelite Member and its stratigraphic position lying above a marble unit and overlain by a thick homogenous quartzite unit (the Leckee Quartzite Formation) strongly suggests that it is a correlative of the Port Askaig Tillite.

2.1.2 Leckee Quartzitic Formation

The Leckee Quartzitic Formation consists dominantly of white to light-grey quartzite, and is the most distinctive and readily identifiable unit in the Ox Mountains. The unit is well exposed along the southeast of the granodiorite between (M 1525 9500) and (G 3970 1465), and in a number of rafts within the granodiorite. It undergoes rapid tectonic thickness variations from approximately 2300m at Lough Anaffrin (M 1675 9615) to 580m at Lough Talt (G 3970 1465), where it forms the core of a major F3 fold. It is generally highly deformed; original sedimentary structures are therefore rare and generally obscure. The only area in which the way up of the unit can be determined with any confidence is a small area north of the Tawneyshane Pelite Member at Lough Anaffrin (M 1685 9641), (Figure 2.5), where heavy mineral bands emphasise lithological banding, enabling cross-laminations to be observed (Plate 2.3–2.4) In this area the sedimentary structures indicate that the beds are downwards facing in S3, that is they young to the northwest, and prior to F3 folding the rocks were inverted and dipping southeast.

Long and Max (1977), cite way up evidence within the Leckee Quartzitic Formation south of the Tawneyshane Pelite Member that indicates that the beds are upwards facing in S3, that is they young away to the southeast. This implies that this unit is exposed within the core of an F2 antiform, as F3 minor structures consistently verge to the northwest in this part of the section. Although the area was mapped in detail during the present study, the present author was unable to locate these examples, hence the stratigraphic succession developed in this chapter, with the Tawneyshane Marble Member at its base, is based on additional criteria such as the bilateral symmetry of the Tawneyshane Pelite Member and regional stratigraphic correlation and not on the three examples of sedimentary way up evidence from northwest of the Tawneyshane Pelite Member, which alone cannot distinguish between the possibility of the presence of an F2 fold and a stratigraphic lens.

SE NW

SE NW

As exposure of the Tawneyshane Formation has a very limited lateral extent, it is extremely difficult to determine the position of the F2 fold axis and hence the location of the base of the Leckee Quartzitic Formation further to the northeast and southwest. At Lough Anaffrin, the basal part of the formation consists dominantly of a flaggy, muscovite rich, white–weathering quartzite, with minor associated bands of psammite and orthoquartzite which rarely exceed 20cm in thickness. Rare cross-laminations are seen where heavy mineral bands have emphasized lithological banding. The formation was regarded previously as entirely composed of quartzite, but several aluminous pelite and semi-pelite bands have been discovered by the present author close to the stratigraphic base of the formation at Callow Loughs (G 3052 0392), where large sillimanite porphyroblasts (maximum length 5cm) are present. These have also been observed at Coolagagh (G 2980 0470), and are the product of the thermal metamorphism resulting from the intrusion of the granodiorite.

Quartzitic grits, containing quartz grains up to 6mm in diameter are rare, and are only observed at (M 1882 9722), (M 1887 9724) and (G 3328 0575), in the upper part of the formation which is generally purer and more massively bedded than the base. Calc-silicate schists, white to pale-green in colour, composed of diopside, actinolite, epidote, biotite and microcline are also concentrated towards the top of the formation and occur east of Crummus (G 3925 1468) and Kilmore (G 2803 0131) where they reach a maximum thickness of 10m. Due to poor exposure it is not possible to trace these calcareous units for any distance along strike.

For most of its exposed length the southeastern boundary of the Leckee Quartzitic Formation is marked by the Lough Talt Slide which separates the upper part of the formation from the Ummoon Formation to the southeast. The upper boundary of the formation is also exposed at Lough Talt at the northern end of the granodiorite, where it lies on the northwest dipping limb of the major F3 fold that forms the core of the inlier. In this area the Lough Talt Slide is absent and the Leckee Quartzitic Formation is separated from the pelites and semi-pelites to the northwest (discussed in detail in area 3), by a unit comprising pelite, semi-pelite, feldspathic psammite, and dark weathering quartzite. This unit which is termed the Leckee Transition Member (section 2.2.1), is not present on the southeast side of Crummus (G 3947 1475), where the Lough Talt Slide (section 4.3.5.1) brings quartzites of the Leckee Quartzitic Formation into abrupt tectonic contact with pelites, semi-pelites and psammites of the Ummoon Formation. It provides critical evidence for the correlation of the rocks exposed on either side of the granodiorite, (areas 1 and 3)

2.1.3 Ummoon Formation

The Ummoon Formation is exposed over a greater area than any of the other stratigraphic units in the Ox Mountains and underlies some 190km2 of the area between Slievenagark Lough (M 1460 9320) in the southwest to its tectonic contact with the granulites of the northeast Ox Mountains to the north at the Ladies Brae.

The unit is dominantly semi-pelitic in character with subsidiary psammite and pelite, and lacks any distinctive features or marker horizons that enable it to be distinguished easily from units in the Ox Mountains such as the Lismoran Formation which also dominantly consists of monotonous banded psammites, semi-pelites and pelites. As a result, some areas have been interpreted as comprising the Lismoran Formation by some workers, and as representing the Ummoon Formation by others.

The Ummoon Formation was originally distinguished from the Lismoran Formation by Taylor (1968), who noted that the former unit was more pelitic in nature. He observed that it was separated from the Lismoran Formation by the

Glennawoo Slide, except for a small area at Toomore (G 2978 0125), where he recorded a normal stratigraphic boundary between the two formations. Phillips et al (1975) also described this boundary as gradational and arbitrarily defined the base of the more psammitic Lismoran Formation as the incoming of psammite bands > 1m in thickness, which approximately coincided with the location of the Glennawoo Slide. Andrews et al. (1978), added to this definition the criterion of garnet size, as they regarded the Lismoran Formation to be characterised by the presence of garnets < 2mm in diameter and the presence of psammite bands > 1m in thickness. They used this criterion to map out a sinuous boundary between the two formations in the poorly exposed area around the headwaters of the Mad, Moy and Owenboy rivers southwest of the Ladies Brae Fault.

The differences between these two formations in the Ox Mountains was considered to be a relatively minor problem until the work of Winchester et al. (1988), who studied the geochemistry of the metavolcanic units within each of the formations. On the basis of this study they regarded the rocks on either side of the Glennawoo Slide as belonging to two different tectono-stratigraphic terranes. They do not regard the great lithological similarity between both units as significant and drew up a further list of four characteristics to distinguish between the two formations (Figure 2.6).

1. While the basic sedimentological characteristics of bed thickness and grainsize are accepted as valid criteria on which to differentiate between the two formations, it should be emphasised that that the definition of the Ummoon Fm as lacking the psammite bands > 1m thick, that characterise the Lismoran Fm is an arbitrary one. There is a gradual decrease in the psammite contact of the metasediments of the Lismoran Fm towards the southeast. This boundary is therefore an arbitrary marker on a gradual lithostratigraphic trend and does not represent a sudden change in sedimentary environment.

FIG 2.6

Table 1. *Lithological characteristics which collectively distinguish between the Ummoon and Lower Lismoran Formations*

Ummoon Formation	Lower Lismoran Formation
1. Interbedded pelite, semi-pelite, psammite, feldspathic semi-pelite, feldspathic psammite, with rare graphitic schist, pebbly grit and quartzite. Dominantly banded semipelitic schist. Beds commonly 5–50 cm thick. No thick psammitic or feldspathic psammite beds over 1 m thick.	1. Generally more psammitic than the Ummoon Formation, with thicker psammite and feldspathic psammite beds than those in the Ummoon Formation, commonly >1 m thick. Rare pebbly grits and graphitic schists.
2. Presence of aluminous pelites with large almandines 1–3 cm in diameter, sometimes with staurolite up to 1 cm long. North-eastwards from Zion Hill pale blue kyanite with whitish edges is also present. West of the Knockaskibbole Fault chloritoid and not staurolite is present. Plagioclase + quartz + white mica + biotite ± almandine ± staurolite ± kyanite is a common assemblage. Kyanite rarely attains lengths exceeding 3 cm, commonly being up to 1 cm long. Pale green micaceous shimmer aggregate commonly has partly replaced kyanite and staurolite.	2. Garnets are <5 mm, and most commonly <2 mm in diameter. Staurolite is apparently absent, except possibly adjacent to the Glennawoo slide zone where it may be difficult through lack of outcrop to locate the slide exactly. Kyanite is apparently absent.
3. Metamorphic grade is staurolite/kyanite (lower) amphibolite facies, except west of the Knockaskibbole fault.	3. Metamorphic grade is believed to be epidote amphibolite facies (quartz–albite–epidote–almandine grade).
4. Basic metavolcanic member (Newantrim Member) near base of formation, close to tectonic contact (Lough Talt slide) with the Leckee Quartzitic Formation.	4. No basic metavolcanic units.

NOTE: Big garnet (+staurolite + kyanite) pelite is generally considered diagnostic of the Ummoon Formation, but its absence may not be taken to identify Lower Lismoran Formation since, as even in well exposed areas, undisputed Ummoon Formation may reveal little or no big garnet pelite.

After Winchester et al. (1988)

2. The occurrence and size of metamorphic index minerals depends on a number of factors, such as reaction and growth kinetics, the bulk rock geochemistry and on the P-T path. In the Ox Mountains the changes in the latter do not coincide with stratigrapic boundaries. This is clearly demonstrated by the difference in the metamorphic grade of the Ummoon Formation which has been subject to greenschist facies metamorphism west of the Knockaskibole Fault, but metamorphosed under amphibolite facies conditions to the east. Therefore this is not considered by the present author to be a valid criterion on which to distinguish between formations.

3. The metamorphic grade of the formations is defined by the presence of various metamorphic index minerals and will therefore be unsuitable for differentiating between formations, for the reasons discussed in 2. above.

4. The statement of Winchester et al. (1988) that the Lismoran Formation can be distinguished from the Ummoon Formation as the former lacks a basic metavolcanic unit is greatly misleading. While it is true that no metavolcanic unit is present within the Lower Lismoran Formation, the Lismoran Formation is divided into an upper and lower formation which are separated by the Callow Member, a major metavolcanic unit.

In summary, the last three points proposed by Winchester et al. (1988) are not considered to represent valid criteria to distinguish between the two formations. The present author has therefore retained the arbitrary division between the two formations originally proposed by Phillips et al. (1975), and considers that the Ummoon Formation passes into the Lismoran Formation via a sedimentary transition and notes that this boundary is an arbitrary one separating stratigraphic units which are very similar.

As the Ummoon Formation on the southeast side of the granodiorite is separated from the Leckee Quarzitic Formation to the northwest by the Lough Talt

Slide, way up criteria from within this quartzite unit cannot be used to infer the way up of the Ummoon Formation. No sedimentary structures have been observed in the Ummoon Formation by previous workers or during the present study. The structurally lowest part of the formation exposed in area 1 outcrops immediately southeast of the Lough Talt Slide at (G 2870 0065), where mylonitised quartzites of the Leckee Quartzitic Formation are brought into contact with the psammites, semi-pelites and pelites of the Ummoon Formation. Pelite and semi-pelite bands predominate, forming 60-70% of the outcrop, while psammite and feldspathic psammite bands generally < 1m thick (average 50cm) make up the remainder of the exposure.

Schistose amphibolite of the Newantrim member (Long 1977) occurs 50-100m southeast of the Lough Talt Slide. The amphibolite which is exposed intermittently along a strike length of approximately 77km, exhibits considerable thickness variation, from 45m east of the Knockaskibole fault (M 1470 9307) to 1m at Callow (G 3135 0225), to approximately 50m at Lough Minnaun (M 5220 2320). Elliptical masses of amphibolitic material up to 25cm in diameter, contained within a matrix of schistose amphibolite consisting of hornblende, plagioclase and epidote, were recorded by Long and Max (1977) at Slievenagark Lough (M 1470 9307). They interpretated these masses as volcanic bombs. Unfortunately the amphibolite facies metamorphism has obliterated any original igneous textures, so the masses therefore lack chilled margins or internal structures that might be used to determine if they represent bombs or pillows. However either possibility is inconsistent with the intrusive origin proposed for the unit by Phillips et al (1975). The geochemistry of this and other metabasalts in the inlier and the implications for terrane accretion models are discussed in chapter 6.

Lithologies present southeast of the Newantrim member are generally less pelitic than those northwest of it. Pebble beds, composed of quartz pebbles up to 4cm in diameter are developed at the Mad River (G 4800 1953) and at Lough

Easky, (G 4557 2050) where they are interbanded with thin graphitic pelite and pelite beds containing large garnet porphyroblasts. The thickness and proportion of psammite and feldspathic psammite bands gradually increases towards the Glennawoo Slide. On the southeast flank of the inlier this major tectonic discontinuity coincides with the boundary of the Ummoon Formation and the Lismoran Formation, arbitarily defined by Phillips (1975) as the incoming of psammite and feldspathic psammite bands more than 1m in width. Detailed mapping conducted during the present study does not support the conclusion of Andrews et al. (1978), that the boundary between the Ummoon and Lismoran Formations follows a sinuous course in the area south of the contact with the granulite facies rocks of the northeast Ox Mountains. The Glennawoo Slide which here separates the two formations, changes its orientation from a northeast to an easterly trajectory, and passes out of the inlier beneath the Carboniferous cover sequence north of Clonacool and Tubercurry, approximately 1km east of the River Moy (G 5350 2157). The area at the upper reaches of the Mad, Moy, and Owenboy rivers is therefore considered to be entirely underlain by rocks of the Ummoon Formation.

For almost all of its outcrop length the Ummoon Formation is separated from the Lower Lismoran Formation exposed to the southeast by the Glennawoo Slide. The single exception to this is on the northern slopes of Knockachree (G 5120 2874), where there is a gradational interbanding of garnet-staurolite semi-pelites and pelites of the Ummoon Formation with the feldspathic psammite, and garnetiferous semi-pelite of the Lower Lismoran Formation. In this area although the rocks are pervasively deformed there is no evidence for the existence of a major tectonic discontinuity or slide. The presence of this transitional contact between the Ummoon Formation and the Lismoran Formation is of critical importance, as it establishes a stratigraphic link between the two formations which Winchester et al (1988) regard as belonging to different tectono-stratigraphic terranes, separated by the Glenawoo Slide which they regard as a major terrane boundary.

Detailed mapping reveals that the pelitic and semi-pelitic rocks exposed on the northwest side of the inlier (Attymass Group of Taylor (1968)) overlies a unit containing a wide range of lithologies (the Leckee Transition Member, section 2.3.1), that is in stratigraphic contact with the Leckee Quartzitic Formation to the SE. The SE margin of the exposure of Leckee Quartzitic Formation, is marked by a syn-metamorphic discontinuity, the Lough Talt Slide. This structure can be traced for 55km SW along strike, and cuts out an unknown amount of stratigraphy, to bring the pelites, semi-pelites and psammites of the Ummoon Formation into contact with the Leckee Quartzitic Formation. The present writer considers that this structure has cut out the Leckee Transition Member and an unknown amount of the base of the Ummoon Formation from the SE flank, thus accounting for the apparent difference in stratigraphy on either limb of the major upright fold whose axis is parallel to strike and contained entirely within the Leckee Quartzitic Fm.

On the northwest side of the inlier, at Knockachree (G 5120 2874), pelites, semi-pelites and psammites of the Ummoon Formation pass upwards via a transition into a more psammitic unit, consisting of pale plagioclase rich psammite with minor feldspathic semi-pelite, originally identified as Leckee Quartzitic Formation by Andrews et al. (1978). The highly feldspathic nature of both the psammitic and semi-pelitic lithologies, and absence of any pure quartzite bands, are consistent with the identification of this upper unit as Lower Lismoran Formation.

2.1.4 Lismoran Formation

The Lismoran Formation, is well exposed along the southeast side of the inlier between Slievenagark Lough (M 1470 9290) and the Mad river (G 4815 1920). In addition, areas of feldspathic psammite and feldspathic semi-pelite at Glennawoo (G 4070 1280) and Knockachree (G 5200 2874), mapped by Andrews et al. (1978) as Leckee Quartzite lacks the pure orthoquartzite that characterise this formation, and have therefore been re-interpreted as Lismoran Formation. A further area of the formation has been identified at Ummoon (G 2900 0005), south of the Glennawoo slide, that was previously regarded as Ummoon Formation by Taylor (1968).

Taylor (1968), working in the central Ox Mountains originally divided the Lismoran formation into an upper and lower unit. It must be stressed that this division is an arbitrary one based on the existence of the Callow Member within the formation: it does not coincide with an abrupt change in the lithological characteristics of the upper and lower units which are extremely difficult to separate on lithological grounds alone. In areas where the Callow Member is not exposed, it is only possible to distinguish the upper and lower parts of the formation by examining a large number of outcrops and assessing the relative proportions of semi-pelite and psammite present, as the upper part of the formation can usually be distinguished from the lower by its slightly more semi-pelitic character.

The Lismoran Formation shows remarkably little lateral variation in lithology throughout its exposure length of at least 70km. The unit has experienced pervasive deformation throughout all of its outcrop, therefore no sedimentary structures have been recorded.

2.1.4.1 Lower Lismoran Formation : The structurally lowest part of the Lismoran Formation is exposed on the northern slopes of Knockachree (G 5120 2874), where it is represented by grey-green garnetiferous, banded semi-pelite (20cm-1m thick), interbanded with bands of psammite and feldspathic psammite (commonly >1m thick), which make up to 60% of the formation. The lateral extent of exposure in this area is limited by extensive peat cover, therefore information on the lithostratigraphy of the remainder of the formation must be obtained from the extensive area of exposure on the southeast side of the granodiorite, southeast of the Glennawoo Slide. At Callow Lough Upper (G 3170 0300), good exposure enables an almost continuous section to be examined through the formation from its tectonic contact with the Ummoon Formation to the northwest, to the Callow Member to the southeast. Here dark-greenish-grey thick bands of feldspathic psammites (commonly >1m thick) dominate forming up to 60% of the outcrop, these are interbanded with garnetiferous feldspathic semi-pelites and pelites, that are occasionally tourmaline bearing and often contain a small amount of opaques. The proportion of pelite and semi-pelite gradually increases towards the contact with the Callow Member to the southeast.

2.1.4.2. Callow Member: The Callow Member is exposed along a strike length of 67km from Shanvolley (G 1650 0101), to Sussuecommon (G 4602 1507), and is a black schistose amphibolite composed of hornblende, plagioclase, epidote, quartz, chlorite and sphene. Irregularly shaped calcite veins and pods of yellowish-green epidote (up to 12cm in diameter) are common and occur in layers parallel to the intense tectonic fabric developed throughout the unit. Phillips et al. (1975) describe a 10m wide zone of increased strain along the southeast side of the member as a D2 tectonic slide (D3, in the present chronology). Detailed mapping has revealed a comparable increase in strain along the northwest contact with the Lower Lismoran Formation (Fig 2.4). The present writer considers that these local increases in strain result from the rheology contrast between the more competent amphibolite and the less competent semi-pelite and psammite of the metasediments. These high strain zones are concordant with lithological banding in the metasediments and do not appear to have cut out any stratigraphy, they are therefore not mapped as significant tectonic breaks by the present author.

Andrews et al. (1978), regard the unit as intrusive in origin and cite cross-cutting relationships with the metasediments of the Lismoran Formation at Suessuecommon (G 452 155), in support of this. This conclusion is not accepted by the present author as no examples of intrusive relationships were observed during the present detailed study of this area. In addition, the Callow Member everywhere appears to be parallel with the lithological banding in the metasediments that form its northwest and southeast boundary. The extrusive origin of the Callow Member is confirmed by the discovery of a number of large elliptically shaped clasts of volcanic material at Callow (G 3374 0430) which range in size between 25 and 50cm in diameter, and are randomly distributed in and supported by a matrix of schistose amphibolite. These clasts are randomly distributed in the matrix and do not occur in parallel trains. In addition their long axes make a moderate angle to the well developed foliation in the matrix. Together these features effectively eliminate the possibility that they represent boudins of an originally coherent body. Similar clasts were recorded in the Newantrim Member by Long and Max (1977), who considered that they were volcanic bombs. The amphibolite facies metamorphism and deformation has totally obliterated the original igneous textures, they therefore lack chilled margins or internal structures that might be used to determine their origin. With the absence of internal textural evidence it is difficult to determine whether the clasts represent bombs or pillow basalts that have subsequently been reworked. While their large size and narrow size distribution suggests that their origin may have been as pillows, they lack the features such as concentric cooling joints and an interlocking or keystone geometry that would make a positive identification possible. The geochemical correlation of this unit and the Newantrim Member with the other Dalradian metavolcanic suites in W Ireland, and its implication for terrane accretion models within the Ox Mountains is discussed in chapter 6.

2.1.4.3 Upper Lismoran Formation: The Upper Lismoran Formation is extremely difficult to distinguish from the lower part of the formation which is lithologically very similar. Both parts of the formation consist of greenish-grey semi-pelite and feldspathic semi-pelite interbanded with thinner bands of psammite and feldpathic psammite. Excellent exposure enables an almost continuous section to be examined through the unit at Lismoran (G 3250 0250), where a gradual decrease in proportion and thickness of psammite and feldspathic psammite bands can be observed towards the southeast. A 40m exposure gap separates the Cloonygowan Formation from the most northeasterly exposure of the Upper Lismoran Formation. The 700m thickness of sediment exposed therefore represents a minimum thickness estimate.

2.2 THE LINK BETWEEN THE STRATIGRAPHY OF THE NW AND SE SIDES OF THE GRANODIORITE : AREA 2

The Ox Mountains Granodiorite and rocks of area 2 separate the Leckee Quartzitic formation of the south east side of the inlier (area 1) from the pelite, semi-pelite and psammite of area 3, (Attymass Group of Taylor (1968), Ellaghmore, Killgellia and Carrick O' Hara and Leckee Quartzitic Formations of Andrews et al. (1978)). The rocks of area 2 are very distinctive and consist of rapidly alternating bands of calc-silicate schist, pelite, semi-pelite, dark psammite, quartzite and schistose amphibolite, here termed the Leckee Transition Member. This member is exposed at Tawnyany (G 3785 1582) and on Benalta (G 3840 1630). This area of exposure provides the critical link between the stratigraphy exposed on either side of the granodiorite (areas 1 and 3), and is the stratigraphic basis for regarding the Attymass Group (Taylor,1968) as a direct equivalent of the Ummoon Formation southeast of the granodiorite.

2.2.1 Leckee Transition Member

Metasediments of the Leckee Transition Member are exposed between (G 3787 1606) and (G 3925 1490) south of the Gap (G 3787 1625) and on Benalta (G 3867 1650) to the north, where they comprise rapidly alternating bands of calc-silicate schist, pelite, semi-pelite, dark psammite, quartzite and schistose amphibolite. The contact of the Transition Member with the structurally underlying Leckee Quartzitic Formation is not exposed; the boundary between the two units is therefore defined arbitrarily by the first occurrance of massive pale quartzite, in the structurally lowest part of the unit, which occurs on Crummus (G 3922 1480). The most complete section through the member is exposed on Benalta (G 3867 1650), where pelite, semi-pelite and minor psammite bands pass up into a 15m thick band of calc-silicate schist that can be traced for 1500m east along strike in gently dipping beds. Further bands of calc-silicate (closely associated with and commonly interbanded with thin bands of amphibolite 1-10cm thick) are present further up the sequence which is dominated by semi-pelite and dark psammite bands which are generally $<$ 5m thick, and are dark grey, in contrast to the very pale psammite of the Leckee Quartzitic Formation.

Thicker amphibolite units are more common towards the structurally highest parts of the unit, which gradually passes upwards into the more homogeneous pelites and semi-pelites of the Ummoon Formation (Attymass Group of Taylor (1968)), exposed to the northwest.

These metasediments have been identified as Attymass Group, Ummoon Formation and Leckee Quartzitic Formation by Taylor (1968), and as Attymass Group and Leckee Quartzitic Formation by Andrews, Phillips and Molloy (1978), both of whom proposed complex structural models (discussed fully in chapter 4) to account for the observed stratigraphy. This interpretation implies extreme facies variation within the Leckee Quartzitic Formation as it requires the highly variable pelite, semi–pelite, psammite, limestone and amphibolite exposed at Tawneyany (G 3785 1582), to be stratigraphic equivalents of the homogenous pure white quartzite of the Leckee Quartzitic formation exposed on Benalta (G 3867 1650).

The interpretation, by the present author, of these metasediments as a transition member or lateral facies variation within the Leckee Quartzitic Formation greatly simplifies the stratigraphy and is consistent with the observed structure. The structural model for this area is discussed fully in chapter 4, where it is shown that vergence data indicate the presence of a major F3 antiformal hinge located within the Leckee Quartzitic Formation exposed on Crummus. On the southeast side of this structure the Lough Talt Slide brings the Ummoon Formation into tectonic contact with the Leckee Quartzitic Formation cutting out the Leckee Transition Member.

2.3 STRATIGRAPHY NORTHWEST OF THE GRANODIORITE, L CONN - TAWNANEILLEEN : AREA 3

A large area of metasediments is exposed on the northwest side of the inlier between the eastern shore of Lough Conn (G 2100 0700) and Knockachree (G 5162 2857). For most of its exposed length Carboniferous cover forms the northwest limit of exposure. However at Lough Brohly (G 3100 1475) rocks of the North Mayo Inlier (Long and Max 1977) are exposed in close proximity to the rocks of the Ox Mountains Succesion to the southeast. The boundary between the units is not exposed. The Granodiorite and areas of blanket peat separate this area from the rocks of area 1 for the entire length of the inlier.

Long and Max (1977), consider that these rocks are equivalent to the Ummoon Formation southeast of the granodiorite, whereas Andrews et al. (1978), consider that they are stratigraphically below the Leckee Quartzitic Formation, which they believe structurally overlies them. The absence of distinct marker horizons makes lithostratigraphic correlation between both areas difficult, however when considered alongside the structural evidence discussed in Chapter 3, the lithological characteristics of the rocks of area 3 described below is consistent with their identification as Ummoon Formation.

2.3.1 Leckee Transition Member (Corradrishy Quartzite after Taylor, (1968))

A thinly banded (generally < 1m thick), greenish grey psammite and quartzite unit, (the Corradrishy Quartzite, Taylor (1968)), is the structurally lowest unit exposed along the northwest flank of the inlier. The Ox Mountains Granodiorite forms the south west limit of its exposure at Roosky Lough (G 2900 0992). The base of the unit is not exposed, therefore the 80m thickness of metasediment exposed represents a minimum thickness for the formation. The proportion and thickness of psammite bands rapidly decreases upwards from 10-30cm, (average 20cm), at north of Corradrishy (G 2789 0906), where they make up to 95% of the exposure, towards the transitional contact with the structurally overlying dark-grey pelites and semi-pelites of the Attymass Group north of Corradrishy (G 2789 0906).

This thin unit was assigned formation status by Taylor (1968), who considered that it represented the base of the Attymass Group. Long and Max (1977) included these metasediments which they termed the Corradrishy Member, within the Cappagh Formation (which they regarded as containing all of the Attymass Group of Taylor), and concluded that their stratigraphic position was uncertain. The present author considers that this unit is most easily explained either as as part of the upper part of the highly variable Leckee Transition Member or a lateral facies transition within the Ummoon Formation, which structually overlies it.

2.3.2 Ummoon Formation (Attymass Group after Taylor, (1968); Ellaghmore, Kilgellia and Carrick O' Hara Formations after Andrews et al. (1978)).

A thick sequence of pelites, semi-pelites and psammites is discontinuously exposed along the northwest side of the inlier, along a strike length of 30km from Illannaglashy (G 2100 0640) in the southwest, to Tawnaneilleen (G 4130 2174) in the northeast. The structurally lowest part of the unit is exposed north of Lough Talt, on the northwest slopes of Benalta (G 3825 1700) and further along strike to Tawnaneilleen to the northeast, where the rapidly alternating sequence of feldspathic pelite, semi-pelite, psammite, calc-silicate and amphibolite of the Leckee Transition Member pass upwards into coarse, pale weathering, muscovite rich pelites with garnet porphyroblasts (maximum dimension 2cm), and staurolite (maximum dimension 5mm). There is no indication that this contact is tectonic.

Minor feldspathic psammite bands are also present however and are normally < 50cm thick and constitute < 20% of the outcrop. Rare calc-silicate lenses, containing diopside, tremolite, clinozoisite and sphene are also present at Tawnaneilleen (G 4130 2174).

At (G 3775 1620), south of the Gap, a small (1.5 × 2.5km) granite body forms the northeast extremity of the Ox Mountains Granodiorite and separates semi-pelites, pelites and calc-silicates of the Leckee Transition Member from garnet-feldspar pelites, semi-pelites and psammites of the Ummoon Formation to the west. The latter lithologies can be traced to Illannaglashy, approximately 20km to the southeast, and also in rafts within the Granodiorite. The contact with the structurally underlying Leckee Transition member is therefore not exposed. The dark, garnetiferous, quartz, biotite, muscovite, chlorite, tourmaline bearing pelites exposed at the structurally lowest levels close to the Granodiorite, rapidly grade up into more semi-pelitic and psammitic lithologies with psammite bands (5-50 cm thick) typically forming 30% of the exposure. The most psammitic lithology present is exposed at Carrowdoogan (G 3130 1200), where psammite bands up to 1m thick make up to 50% of the exposure. Structurally above this the metasediments become more pelitic and tourmaline rich. The structurally highest parts of the formation are exposed closest to the boundary with the Carboniferous rocks to the northwest, where garnetiferous pelites and semi-pelites again dominate the exposure forming approximately 50% of the outcrop. The contact with the overlying Lismoran Formation is not exposed on the northwestern side of the granodiorite.

2.4 STRATIGRAPHY OF SHANVOLLEY, WEST OF THE KNOCKASKIBOLE FAULT: AREA 4

Area 4, in the southwest Ox Mountains is separated from the rest of the inlier by the Knockaskibole fault. The metasediments exposed within the area are highly deformed and the original stratigraphic contacts between units have been obliterated and are now expressed as tectonic slides. Way up criteria have not been observed. The presence of marker horizons; the Leckee Quartzitic Formation, the Newantrim and Callow Members indicates that the stratigraphy of the area represents a continuation of that exposed in area 1, east of the Knockaskibole fault.

2.4.1 The Leckee Quartzitic Formation

The Leckee Quartzitic Formation is the structurally lowest stratigraphic unit exposed in the area. It shows no significant change in lithology from its type locality in area 1, and is everywhere a white weathering, homogeneous, quartzite.

2.4.2 The Ummoon Formation

The Lough Talt Slide separates the Leckee Quartzitic Formation from structurally overlying psammites, semi–pelites and pelites of the Ummoon Formation, in which the Newantrim Member, a thin metavolcanic unit is sporadically exposed.

2.4.3 Lismoran Formation

Psammite, semi–pelite and pelite of the Lower Lismoran Formation is in tectonic contact with the Ummoon Formation. The Callow Member is sporadically exposed close to the structurally highest part of the formation, which indicates that only the lower part of the Upper Lismoran is represented.

2.5 IDENTITY OF THE METASEDIMENTS OF TULLYCOMMONS: AREA 5

The metasediments of area 5 are in fault contact with those of the Ox Mountains Succession of area 4 to the north, and the Cloonygowan Formation of area

6b to the east. Carboniferous rocks onlap to the south and southwest, while the contact with the rocks of the North Mayo Inlier that forms the northwest margin of the area, is not exposed.

Exposure within the area is very poor and restricted to a number of small, isolated outcrops. The metasediments comprise psammites, semi–pelites and pelites interbedded on a scale of 10cm–1m. Semi–pelite and pelite predominate over psammite which forms $< 40\%$ of the exposure. The absence of lithostratigraphic marker horizons precludes positive identification of the stratigraphy. It does however have lithostratigraphic similarities with the banded psammite, semi–pelite and pelite of the Lismoran Formation, and also exhibits an similar tectonic contact to that between the Cloonygowan Formation and the Lismoran Formation at Callow. The high proportion of semi–pelite and pelite to psammite observed in the area suggests correlation with the upper part of the Lismoran Formation. It is also possible that these metasediments represent a higher stratigraphic level of the Lismoran Formation than exposed in area 1.

2.6 CLOONYGOWAN and ARDVARNEY : AREAS 6a and 6b (The Cloonygowan Formation)

The lower greenschist facies metasediments of the Cloonygowan Formation were originally described by Taylor (1968) at Cloonygowan (G 3300 0200). Long and Max (1975) described similar lower greenschist facies rocks at Ardvarney, (M 1525 9500) which they named the Ardvarney Formation. These metasediments clearly represent part of the same stratigraphic unit and are therefore included in the Cloonygowan Formation, the term Ardvarney Formation being discontinued by the present writer.

The contact with the Ox Mountains succession is everywhere tectonic the base of the formation therefore is not observed. The stratigraphic relationship of the lower greenschist facies rocks of the Cloonygowan Formation to the higher grade Ox Mountains Succession is unknown. At Cloonygowan and Ardvarney an exposure gap of a minimum of 25m separates the two units which are lithologically distinct. The conclusion of Phillips et al. (1975), that the boundary between the two units is a 'gradational and probably interbanded one' is therefore not accepted. The top of the formation is not exposed at the present exposure level at Ardvarney and is concealed by Carboniferous rocks at Cloonygowan. The 750m thickness of metasediment exposed in the Clydagh River between Conloon (M 9591 1500) and Ardvarney (M 1525 9500) is therefore a minimum thickness.

Three lithotypes can be distinguished.

1. Light greenish-grey phyllites, composed of muscovite, chlorite, quartz, albite and pyrite.
2. Greywackes composed of 5-10% pebbles of plagioclase and commonly opalescent quartz set in a fine-grained matrix of chlorite, quartz, plagioclase and pyrite.
3. Coarse Greywackes, consisting of 25-60% pebbles of maximum diameter 4cm set in a fine grained matrix of chlorite and quartz.

Beds vary in thickness between 0.2m and 3.0m (average 1m). They are poorly sorted and consist of a wide variety of grain sizes. The thicker beds have a very coarse base with pebbles of vein quartz, feldspar and pelite fragments with a maximum diameter of 4cm. These show an overall grading from base to top, most clearly seen in the Clydagh River (M 1523 9540). Generally smaller grains of blue opalescent quartz are common in the medium grained beds. Purplish phyllite beds reach a maximum thickness of 12m at (M 1523 9530).

The isolated outcrop of amphibolite at (M 1433 9367) that was described by Long and Max (1977) as part of the Ardvarney Formation is considered to represent part of the basic lithologies present along the Knockaskibole fault. No other amphibolite has been recorded from the Cloonygowan Formation.

The poorly sorted nature of the sediment suggests that deposition was rapid, and is consistent with a turbiditic origin for the formation. The sedimentological environment and stratigraphic position of the Clooygowan Formation is discussed in chapter 3.

2.7 THE OX MOUNTAINS IGNEOUS COMPLEX

The Ox Mountains Igneous Complex comprising the Ox Mountains Granodiorite, Lough Talt and Lough Easky adamellites was intruded in two main phases. Intrusion of the Ox Mountains Granodiorite, which forms the core of the inlier at 478 ± 12 Ma., Pankhurst et al. (1976), Max and Long (1977), was followed by the intrusion of the adamellites, dated by Long, Max and O'Connor (1984) at 401 ± 33 Ma.

2.7.1 Ox Mountains Granodiorite

The Ox Mountains Granodiorite is exposed in the centre of the inlier along a strike length of 29km, between Crumlin (M 1660 9765) in the southwest and the Gap (G 3787 1625) in the northeast. It is a maximum of 6km wide and hence is elongate in shape along a northeast-southwest axis, parallel to the regional strike. Exposure is good on the high ground northeast of Foxford and excellent along the clean glaciated surfaces around the shore of Lough Cullin and the south shore of Lough Conn. However exposure is very poor in the area southeast of Lough Talt, where large areas of blanket peat are developed on the gentle upper slopes of the Ox Mountains.

The Ox Mountains Granodiorite comprises four main lithotypes, of which granodiorite forms the greatest part. The intrusive history of the igneous complex has been determined (McCaffrey pers comm.) by examining cross-cutting sheeting relationships. This suggests that intrusion of granodiorite and appinite was preceeded by the intrusion of granitic material and suceeded by the intrusion of tonalite.

Granite is concentrated on the northwest side of the intrusion. Large (approximately 1 × 2 km) oblate bodies of granite are exposed at (G 3760 1626) west of Lough Talt and at (G 2418 0691) southeast of Tawnnaghman where it is generally white to pale pink in colour and composed of medium grained alkali feldspar, quartz, biotite and plagioclase. Muscovite is commonly present at its contacts with other lithologies. A narrow zone consisting of sheets of granite generally $<$ 5m thick that are intruded sub–parallel to the regional S3 fabric in the metasediments cut across early F3 folds, but are themselves highly deformed by continued D3 transcurrent deformation. This indicates that the granite is syn-kinematic with respect to the regional D3 event.

Granodiorite which is coarse grained and grey to dark grey in appearance and composed of alkali feldspar megacrysts (up to 3cm in length) in a groundmass of plagioclase, biotite and quartz forms the largest component of the complex and accounts for $>$ 95% of its exposed area. The contact with the metasediments is discontinuously exposed form Illannaglashy (G 2075 0608) in Lough Conn to Bunnyconnellan East (G 3625 1627) in the northeast where it dips at approximately 45° northwest. The granodiorite intrudes the large granite sheets along this northwestern margin and therefore is considered to slightly postdate its intrusion.

Along its northeastern boundary, west of Lough Talt, the granodiorite is brought into contact with metasediments to the northeast by a minor northwest-southeast trending fault that runs through the Gap. A 400m offset of a thin amphibolite band demonstrates that the horizontal displacement on this structure is dextral. Its vertical displacement cannot be measured, however is unlikely to be large as the metasediments are lithologically very similar on both sides of the structure. Further to the southeast the contact with the Lough Talt Adamellite is concealed beneath extensive peat cover.

The southeast boundary of the granodiorite is a steeply inclined and intensely deformed sheeted contact zone. Sheets of granodiorite generally < 20m wide, intrude metasediments of the Leckee Quartzitic Formation and the Ummoon Formation for several hundred metres southeast of the continuous body of granodioritic material. This contact zone is discontinuously exposed along the southeast side of the inlier from Attimachugh (G 3240 0620) to Sallagher (M 1670 9860), and is best exposed on a small hill west of Callow Loughs (G 3050 0396) and at (G 0357 0535), along the Foxford–Swinford road. The contact itself has been mapped where the proportion of metasediments exceeds 50% of the exposure.

The Knockaskibole Fault forms the western boundary of the granodiorite between Sallagher and Greenans (G 1680 0013), where the fault swings northeastwards to link with the North Ox Mountains Fault and associated high strain zones that are located along the northwestern side of the granodiorite (Figure 2.4). This isolates an 8km^2 area of granodiorite in a compressional bend in the shear zone, and preserves the original shallowly west dipping intrusive contact between the granodiorite and the Leckee Quartzitic Formation on Farbreiga (G 1692 0298).

Intrusion of tonalite is confined to the north west part of the granodiorite where it is a strongly foliated dark grey medium grained rock of uniform texture composed of plagioclase, biotite and quartz. It intrudes granite on a scale of 0.5–2.0 metres at Pontoon (G 2150 0455), where sinistral syn-magmatic extensional shears deform both lithotypes. Small elongate bodies of tonalite also intrude granodiorite southeast of Lough Cullin, therefore the intrusive sequence, granite-granodiorite-tonalite can be derived, (Mc Caffrey pers comm. 1988).

Occurance of appinitic lithologies which dominantly consist of hornblende megacrysts within poikilitic amphibole, is confined to the south east of the inlier, which coincides with the area of granodiorite sheeting in the metasediments and most intense transcurrent deformation. The appinites generally form highly elongate northeast-southwest trending lenses which are up to 4.5km long and 0.5km wide. The largest of these bodies is exposed at (M 1700 9486), south of Lough Anaffrin, where it dominantly consists of hornblendite megacrysts within poikilitic amphibole. Diorites, quartz-diorites and dioritic hornblende also occur, and exhibit complex veining relationships with the appinites. Appinitic lithologies also intrude the granodiorite. At Curranara (G 2830 0308), close to the centre of the granodiorite pluton, intricate magma mixing relationships between appinites and granodiorite indicate that they were emplaced contemporaneously.

2.5.2 The Lough Easky and Lough Talt Adamellites

The Lough Easky pluton is small in size (6km × 2.5km), and located west of Lough Easky (G 4450 2300), approximately 10km north of the northern limit of exposure of the Ox Mountains Granodiorite. It is poorly exposed and therefore its lobate shape has been determined by mapping in the stream sections that cut through the blanket peat cover concealing most of the pluton in the high ground east of Lough Easky. The pluton is best exposed on the gently sloping hillside east of the Lough, where it is a coarse grained adamellite consisting of quartz, oligoclase, microcline (often megacrystic) and biotite. Neither lateral variations in igneous lithology nor xenoliths have been recorded within the pluton. Deformation within the pluton is strongest at its margins and is best observed at (G 4487 2300), 100m east of the lough, where the greenschist facies sinistral deformation fabrics, consisting of an alignment of ductilely deformed quartz ribbons and fractured feldspar megacrysts are overprinted by later more discrete and relatively weak dextral shears.

The Lough Talt Adamellite is located 4km south of Lough Talt (G 3980 1520) where it is in steep tectonic contact with the Ummoon Formation to the southeast.

The northeast contact with the metasediments is exposed at (G 4042 1345), where it dips gently to the southwest. Original intrusive relationships have been discovered here by the present author suggesting that this boundary may represent the base of the pluton. The northwest and southwest contacts of the intrusion are concealed beneath extensive blanket peat; its exact size and shape are therefore unknown. Exposure of the adamellite is best on the steep north facing cliff at (G 4042 1360) south of Lough Talt. Although similar in composition to the Lough Easky Adamellite the Lough Talt adamellite contains a large number of microdiorite xenoliths which are flattened into the plane of the moderately inclined, southeast dipping sinistral S-C deformation fabric that is developed throughout its exposure. This fabric intensifies towards the contact with the Ummoon Formation to the southeast, and as with the Lough Easky Adamellite is succeeded by discrete dextral shears. The emplacement and deformation of all of the Igneous Complex is discussed in more detail in chapters 4 and 5.

2.8 SERPENTINITES

Serpentinite bodies derived from basic and ultrabasic rocks occur commonly along or south of the Fair Head-Clew Bay Line in Ireland, Max and Riddihough (1975), Lemon (1966), Phillips et al (1975). They were first described in the Ox mountains by Currall (1964), who mapped several small (maximum size 10 × 10m), heavily weathered serpentinite bodies along the line of the Knockaskibole fault between (M 1460 9355) and (M 1345 9182). Max (1988) links these bodies, which dominantly consist of dolomite and fibrous crysotile with minor antigorite, sericite and quartz to the serpentinites that occur along the South Clew Bay Imbricate Zone (SCBIZ), where serpentinised dunite-harzburgite breccias, schistose carbonated peridotites and other basic and ultrabasic igneous rocks are in tectonic contact with the Clew Bay Group and Deer Park Schists. Ryan et al (1983), analysed sheeted dykes and serpentinites from the SCBIZ, and concluded that they have the geochemical characteristics of ophiolitic material. If this conclusion is valid and the serpentites of the southwest Ox Mountains share a common origin with those along the south side of Clew Bay, then the Knockaskibole Fault clearly has a greater significance than has previously been recognised. No occurrence of the other lithologic elements, deep sea-radiolarian chert and spilitic basalts and diabases, that characterise ophiolites have been observed in the Ox Mountains either during the present study or by previous workers. This conclusion cannot therefore be confirmed. There is no need to regard this fault which anastamoses with the Clew bay fault Zone as a cryptic suture, as the serpentinite bodies could easily represent smears of the larger bodies exposed on the south side of Clew bay which have simply been caught up in the Knockaskibole fault.

2.9 CONCLUSIONS : The Ox Mountains tectono-stratigraphic Sequence

Lithostratigraphic data from the Ox Mountains have been used to develop a tectono-stratigraphic sequence that can be applied throughout the area (Figure 2.3). This sequence is dismembered by tectonic slides, therefore the way up criteria observed in the Leckee Quartzitic Formation and the Cloonygowan Formation cannot be extrapolated into the rest of the sequence. However, despite the presence of tectonic slides, the identification of stratigraphic transitions between some units permits a stratigraphic sequence to be established.

Way up criteria within the Leckee Quartzitic Formation indicates that this unit is stratigraphically above the Tawneyshane Pelite Member. Identification of the highly variable Leckee Transition Member, that represents the transition between the Quartzite and the Ummoon Formation, establishes an important stratigraphic link between the two formations, which are normally separated by the Lough Talt Slide. This also establishes that the Ummoon Formation is a direct lateral equivalent of the pelite, semi-pelite and psammite exposed on the northwestern side of the granodiorite (Attymass Group after Taylor, (1968); Ellaghmore, Killgellia and Carrick O'Formations after Andrews et al. (1978)), that structurally overlies the Leckee Transition Member.

The criteria cited by Winchester et al. (1988), for distinguishing the Ummoon Formation from the Lismoran Formation are not accepted. The present author has retained the arbitrary division between the two formations proposed by Phillips et al. (1975), and considers that the Ummoon Formation passes into the Lismoran Formation via a sedimentary transition and notes that this boundary is an arbitrary one defined on the thickness of psammite bands, separating units that are very similar. There is therefore no stratigraphic basis for regarding the Glennawoo Slide as a major terrane boundary.

The discovery of large elliptically shaped volcanic clasts within the Callow Member, following the identification of volcanic bombs within the Newantrim Member (Long and Max, 1977), confirms that both units are of volcanic origin.

The stratigraphic relationship of the greenschist facies rocks of the Cloonygowan Formation to the higher grade rocks of the Ox Mountains Succession is unknown. At Cloonygowan and Ardvarney tectonic contacts separate the two units which are lithologically distinct. The conclusion of Phillips et al. (1975), that the boundary between them is a gradational and probably interbanded one is therefore not accepted by the present author.

In the chapter that follows this tectono–stratigraphic sequence is correlated on a formation to formation basis with the rocks of south Donegal which represent the closest area of autochthonous Dalradian, and the sedimentary environment and possible areas of sediment provenance are briefly examined.

CHAPTER 3

REGIONAL STRATIGRAPHIC CORRELATIONS, SEDIMENTARY ENVIRONMENT AND PROVENANCE

3.1 REGIONAL STRATIGRAPHIC CORRELATIONS

The correlation of the metasediments of the central and southwest Ox Mountains (the Ox Mountains sequence developed in chapter 2) with the Dalradian Supergroup originally proposed by Taylor (1969), and developed by Long and Max (1977), is strongly supported by the present correlation with S Donegal, (Fig 3.1), which is the closest area of undisputed autochthonous Dalradian to the Ox Mountains (Hutton and Dewey, 1986). This formation by formation correlation with southern Donegal where the way up and stratigraphy are known suggests that the tectono-stratigraphic sequence developed in chapter 2 also represents the stratigraphic *succession*. This however is not accepted by Phillips et al. (1975, 1977) who regard the rocks of the central, southwest and northeast Ox Mountains as Pre-Caledonian in age and probably as representing the Moine. The lithological basis for this model is examined critically in section 3.1.3, and the principal reasons why it is considered by the present author to be invalid are also discussed.

3.1.1 Correlation of the Ox Mountains Sequence with southern Donegal and other areas of the Dalradian

In this section a formation by formation correlation with southern Donegal will be examined in detail. The regional correlation with the rest of the Dalradian of the British Isles is summarised in Figure 3.2.

FIG 3.1

Lithostratigraphic correlation with southern Donegal

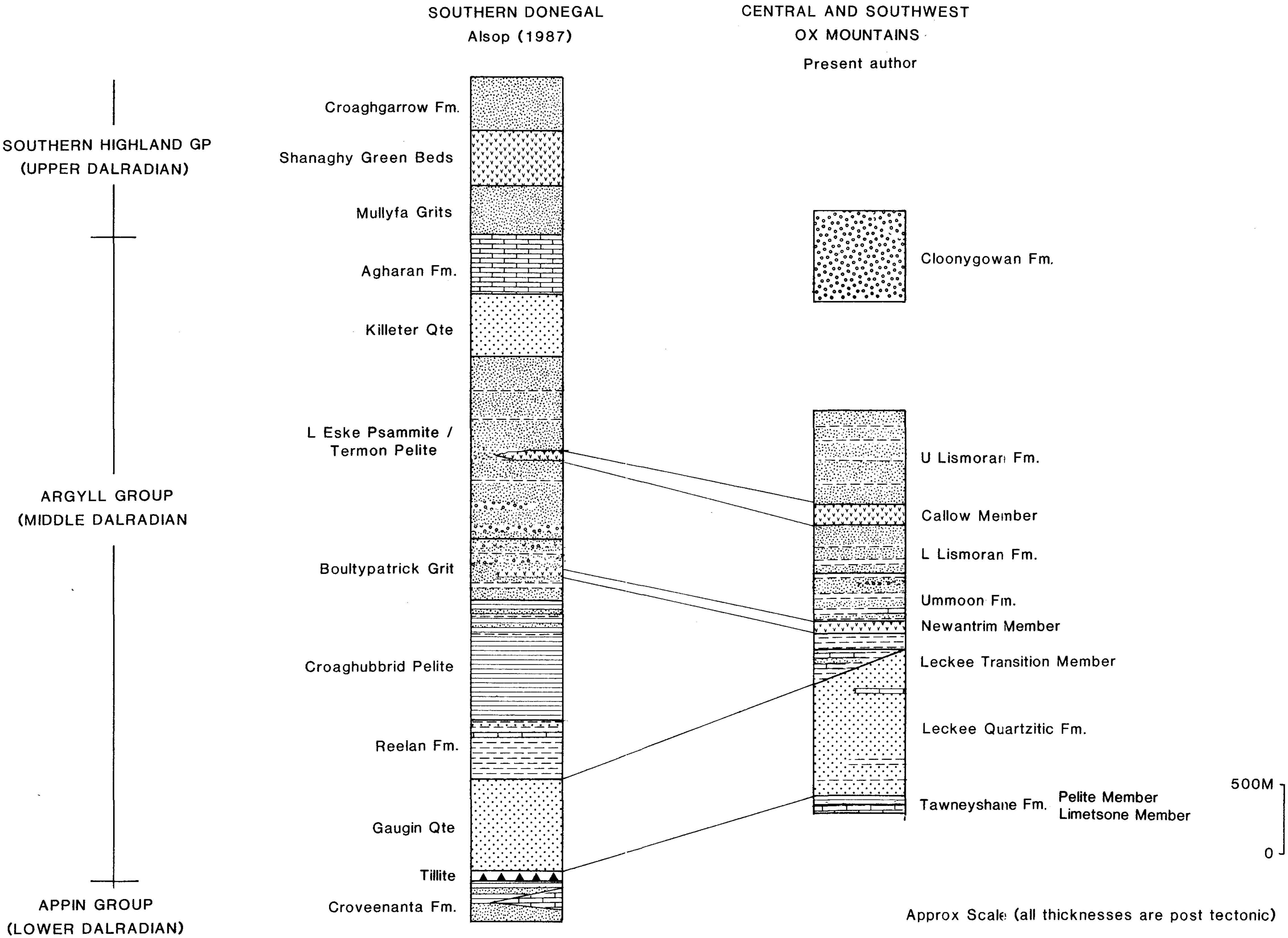


FIGURE 3.2

DALARDIAN SUPERGROUP

Groups	Subgroups	C AND SW OX MOUNTAINS	S DONEGAL (Pitcher, Shackleton & Wood 1971, Pitcher & Berger 1972) Alsop (1987)	GLENCOLUMBKILLE (Anderson 1953, Howarth 1966, Kemp 1966, Pitcher & Berger 1972)	INISHOWEN-FANAD (McCallien 1935, Spencer 1971, Gower 1973, Pitcher & Berger 1972, Roberts 1973)	CONNEMARA (Edmunds & Thomas 1966, Cruse and Leake 1968, Leggo et al 1969, Cobbing 1969)	SCHICHALLION LOUGH TAY KINLOCHLAGGAN CALLANDER (Bailey & McCallien 1937, Anderson 1947, Elles 1930, Rast 1958, Treagus 1969, Harris 1962)	KINTYRE-ARRAN EASDALE-APPIN (Anderson 1947, Bailey 1960, Roberts 1966, Litherland 1970)
Southern Highland (Upper Dalradian)		Cloonygowan Fm.	L Derg Slide Croaghgarrow Shanaghy Green Beds Mullyfa Grits		Greencastle Green Beds Inishowen Head Cloghan Green Beds Fahan Grit Fahan Slate	Benlevy Group	Highland Border Series U Leny Grit Leny Lst & Schist L Leny Grit Aberfoyle Slate Ben Ledi Grit Green Beds Pitlochry Schists	Highland Border Series Lst (? Leny Lst) L Ranza Slates Grits with a lst Green Beds Glen Slaun Schists
Argyll (Middle Dalradian)	Tayvallich		Aghyaran		Culdaff Lst	COrnamona Group	L Tay Lst (some Volc)	Tayvallich Lst
	Crinan		Killeter Qte		U Crana Qte L Crana Qte	Kylemore Group	Ben Lui Schists	Stonefield Schists Crinan Grit
	Easdale	U Lismoran Fm. Callow Member L Lismoran Fm. Ummoon Fm Newantrim Member	Termon Pelite L Eske Psam Boultypatrick Grit Croaghubbrid Pelite	Teelin Point Slieve League	Stragill Linsfort Black Schist Glengad Schist Lag Lst	Lakes marble Gp	Sron Bheag Schist Ben Lawers Schist Ben Eagach Schist Carn Mairg Qte	Craignish Phyllite Easdale Slate
	Islay	Leckee Transition Member Leckee Quartzite Tawneyshane Fm.	Reelan Fm. Gaugin Qte Tillite	Slieve Tooey Qte Kiltyfanned Schist Tillite	Fanad Qte U Dol Tillite	Streamstown Gp Bennabeola Qte Cleggan B B	Killiecrankie Schist Schichallion Qte Dol beds w Boulder Bed Schichallion B B	Islay Qte Dol beds Tillite
Appin (Lower Dalradian)	Blair Atholl		Croveenanta Fm.	Glencolumnkille Dol Glencolumbkille Schist	Fanad Lst Rosakill Schist	Connemara Marble	Pale Lst Banded Group Dark Lst Dark Schist	Lismore Lst Cuil Bay Slate
	Ballachulish							Appin Phylite & Lst Appin Qte Ballachulish Slate Ballachulish Lst
	Lochaber		Slide					Leven Schist Glencoe Qte Binnein Schist Binnein Qte Ellde Schists Eilde Qte

Eilde Flags (Moine)

CORRELATION OF THE DALRADIAN SUCCESSION OF SCOTLAND AND IRELAND

THE PORT ASKAIG TILLITE FORMATION : The Port Askaig Tillite (Kilburn et al., 1964; Spencer, 1969; Howarth, 1971) is recognised over 700km of outcrop of the Dalradian of Scotland and Ireland and maintains its overall characteristics despite changes in the number of tillite beds and thickness of the formation. The stratigraphy of the boulder bed has therefore been determined in great detail (Kilburn et al., 1965; Spencer et al., 1971). It is necessary to examine briefly the sub-division of the unit and its position within the Dalradian Succession before attempting any correlation, (Figures 3.3, 3.4).

THE ISLAY LIMESTONE SERIES: The boulder bed is everywhere underlain by the Islay Limestone series. In the Garvellachs, Islay and Fanad the series essentially consists of shallow water limestone and dolomite. An upper white, cream or pink limestone or dolomite can often be separated from a lower grey limestone, and thin interbedded grits, sandstone and silt beds are common. In Perthshire, the Slieve League Peninsula and Connemara, the Islay Limestone Series includes more pelite and semi-pelite.

THE BOULDER BED: There is a marked overall change in lithology upwards in the formation. Both mixtites and interbeds are dolomite rich at the base of the formation and change to quartzo–feldspathic towards the top, whilst bedded dolomites are only present in the lower half of the formation. The stone content of the mixtites also changes; the lowest contain only intrabasinal stones, whilst extrabasinal stones become more abundant upwards. Kilburn et al. (1965) subdivided the Boulder Bed into four units.

1. The Lower Boulder Bed, characterised by a pale buff dolomitic mixtite containing buff-weathering dolomite clasts (granite clasts are rare < 1%).

2. The Middle Boulder Bed, distinguished from the lower by its more sandy nature, and total absence of dolomite from its clasts and matrix. It is characterised by the presence of a dark green / greenish grey semi-pelitic matrix and more

FIG 3.3

Type Succession of the Port Askaig Tillite (Spencer et al. 1971)

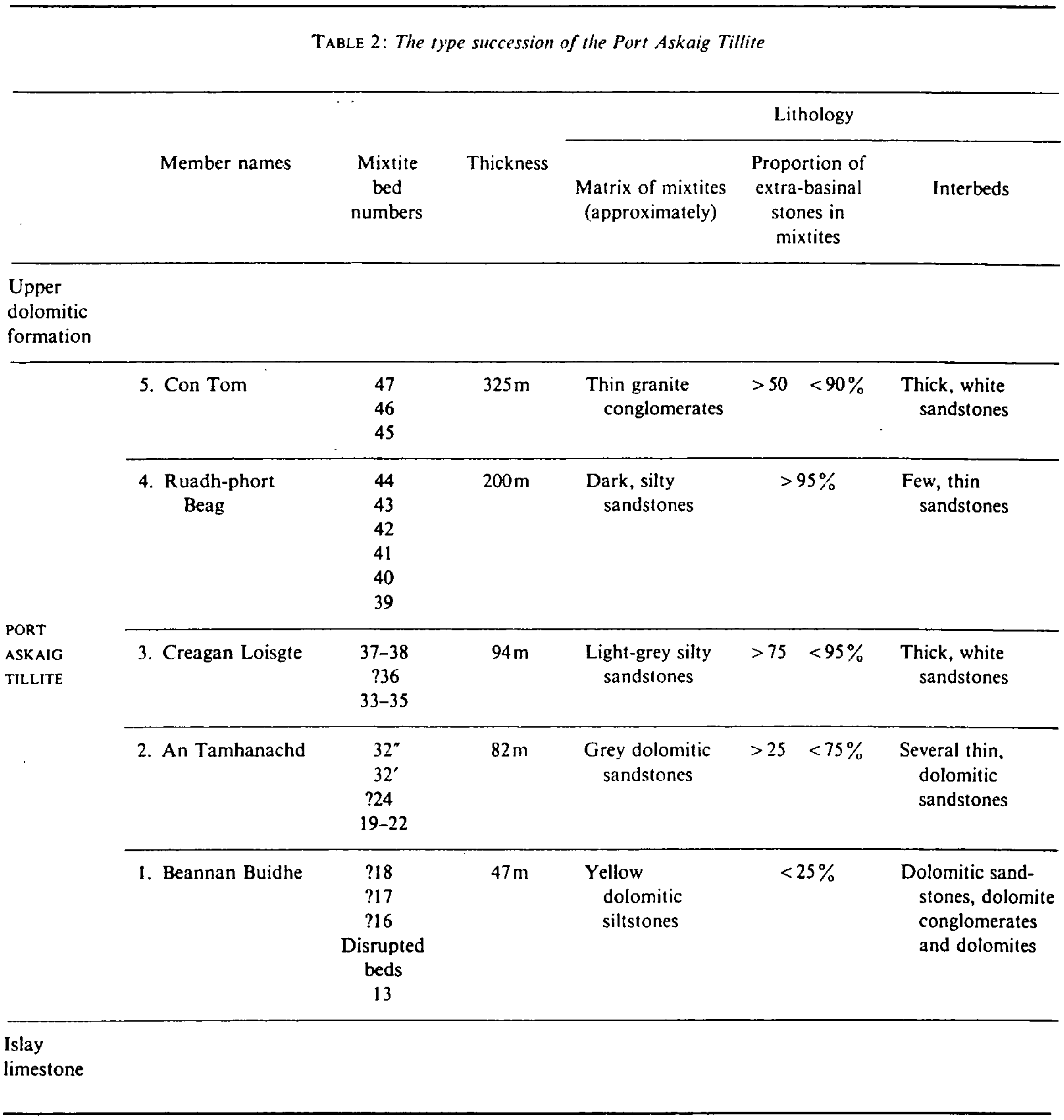
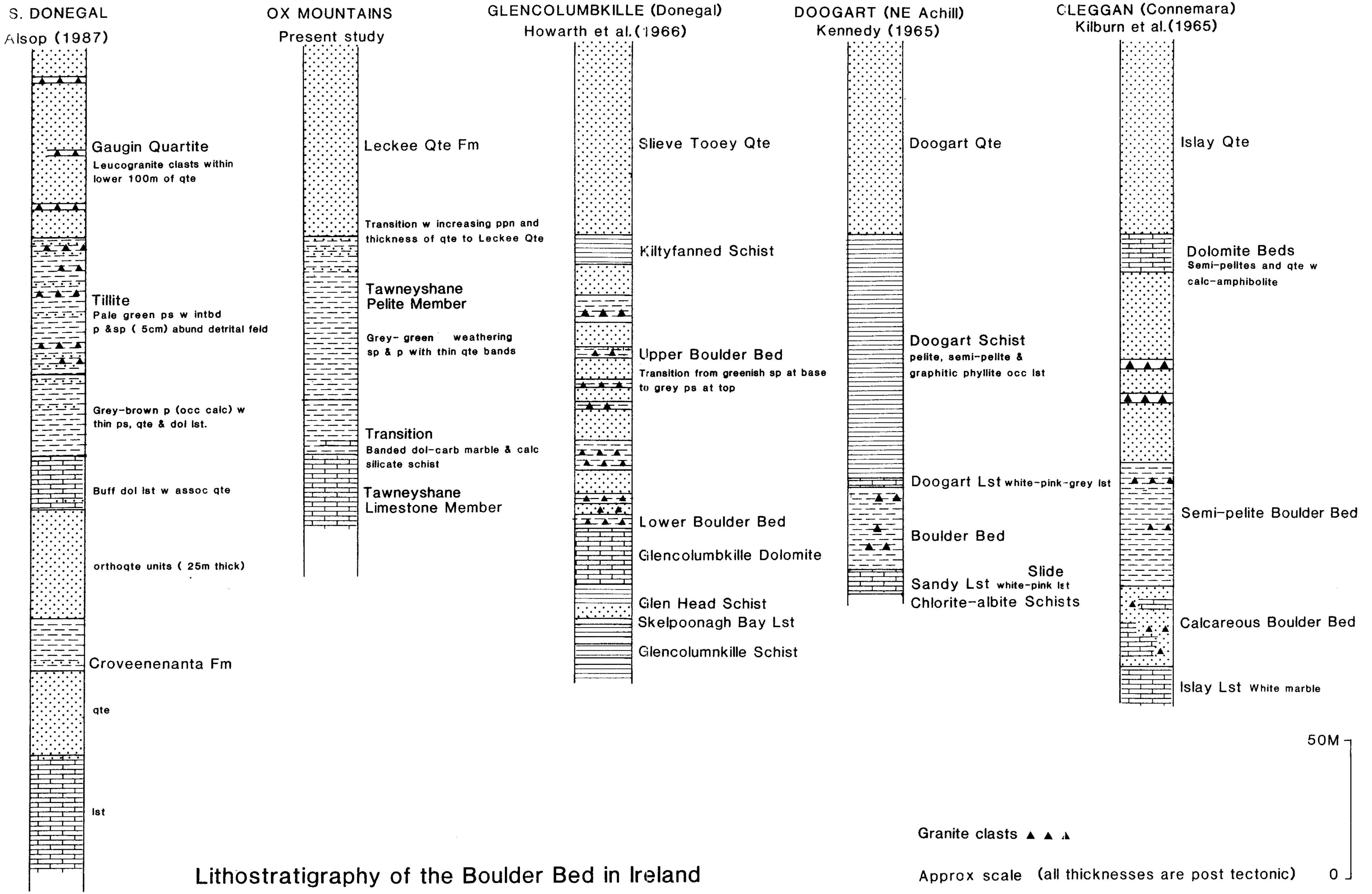


TABLE 2: *The type succession of the Port Askaig Tillite*

				Lithology		
	Member names	Mixtite bed numbers	Thickness	Matrix of mixtites (approximately)	Proportion of extra-basinal stones in mixtites	Interbeds
Upper dolomitic formation						
PORT ASKAIG TILLITE	5. Con Tom	47 46 45	325 m	Thin granite conglomerates	>50 <90%	Thick, white sandstones
	4. Ruadh-phort Beag	44 43 42 41 40 39	200 m	Dark, silty sandstones	>95%	Few, thin sandstones
	3. Creagan Loisgte	37–38 ?36 33–35	94 m	Light-grey silty sandstones	>75 <95%	Thick, white sandstones
	2. An Tamhanachd	32″ 32′ ?24 19–22	82 m	Grey dolomitic sandstones	>25 <75%	Several thin, dolomitic sandstones
	1. Beannan Buidhe	?18 ?17 ?16 Disrupted beds 13	47 m	Yellow dolomitic siltstones	<25%	Dolomitic sandstones, dolomite conglomerates and dolomites
Islay limestone						

FIG 3.4



Lithostratigraphy of the Boulder Bed in Ireland abundant granite clasts, in mixtites that are separated by quartz interbeds < 10m thick.

3. Grey Quartzite with the Upper Boulder Beds. The Upper Boulder beds are composed of pale grey highly feldspathic mixtite with a matrix approaching feldspathic quartzite in composition. Clasts have a high proportion of feldspar. The boundary between units 2 and 3 was defined as the point where an abrupt increase in the thickness and proportion of quartzite intercalations occurs.

4. Upper Dolomitic Formation. The Upper Dolomitic Formation is a laterally variable unit that consists of semi-pelite and quartzite at Cleggan (Connemara) and Fanad (Donegal), but includes calcareous and dolomitic horizons elsewhere.

Spencer et al. (1971) developed this classification further, and divided the Port Askaig Tillite into 5 members from its type locality at Port Askaig, and the surrounding areas (Figure 3.3). These workers defined the base of the formation beneath the lowest mixtite bed and the top above the level of the highest horizon in which they observed extrabasinal stones. They do not regard the Upper Dolomitic Formation (Unit 4, Kilburn et al. 1965) as part of the Tillite. Alsop (1987) records extrabasinal clasts occurring above this stratigraphic level in S. Donegal, the present author therefore prefers to regard this unit as part of the Tillite. The terminology of Kilburn et al. (1965) is therefore retained here.

The Tillite is everywhere overlain by the Islay Quartzite and its equivalents (Figure 3.2), which form one of the most persistent stratigraphic horizons in the British Isles. At its type locality the Islay Quartzite has been divided into a coarse and three fine facies. The former comprises cross-bedded sands and some laminated sand and silt horizons, interpreted as the deposits of shallow marine tidal dunes and other bedforms. These units range in thickness from a few cm to several m. The finer facies comprise laterally persistent, parallel and cross-laminated sand units from mm to a few cm thick.

The lateral variation in lithotype that occurs within the boulder bed and associated horizons in Ireland (Figure 3.4), precludes a member by member correlation to be made with the type locality at Port Askaig in Scotland. However despite local variations in lithology of the beds that overly the Tillite horizons, the boulder bed has three highly distinctive characteristics which enables it to be identified with confidence even where it has been highly deformed.

The most distinctive feature of the Tillite is the presence of extra-basinal clasts, principally of granitic composition. While the occurrence of these clasts positively identifies the Tillite, it must be noted that a large proportion of the formation lacks extra-basinal clasts (Alsop, pers comm.). The absence of such clasts from the Tawneyshane Formation of the Ox Mountains does not therefore preclude correlation of this unit with the Boulder Bed.

While the extra-basinal clasts are the most easily recognised feature of the Tillite, the matrix of the clasts is highly distinctive. Alsop (1987) records the extra-basinal clasts as occurring in a matrix of pale-green psammite with semi-pelite and pelite interbeds. This unit has been examined in southern Donegal by the present author who notes that it is indistinguishable from the psammite, semi-pelite and pelite of the Tawneyshane Pelite Member in the Ox Mountains.

The most compelling evidence for the correlation of the Tawneyshane Pelite Member in the Ox Mountains and the Port Askaig Tillite is the similarity of stratigraphic position of the two units. In all occurrances of the Tillite in Ireland, it overlies a limestone or dolomitic limestone unit and is itself overlain by a thick homogenous quartzite unit (Figure 3.4). The Tawneyshane Pelite Member occupies an identical stratigraphic position in the Ox Mountains, overlying the dolomitic limestone and calc-silicate schist of the Tawneyshane Marble Member and overlain by the thick homogeneous quartzite of the Leckee Quartzitic Formation. The proximity of the Tawneyshane Pelitic Member to the Leckee Quartzitic Formation suggests a correlation with the upper part of the Tillite.

If the upper part of the Tillite in Ireland is examined in detail, it is apparent that it is highly variable and that extra–basinal clasts occur at a variety of stratigraphic levels. Alsop (1987) demonstrated that in southern Donegal pockets of tillite occur within the lower 100m of the Gaugin Quartzite, above the uppermost dolomite / calc–silicate unit (probable Upper Dolomite Formation equivalent). In Glencolumbkille, extra–basinal clasts occur in the transitional unit below the Glencolumbkille Dolomite and the Slieve Tooey Quartzite. In a situation that may be analogous to that in the Ox Mountains, extra–basinal clasts are absent from this stratigraphic level at Doogart on Achill Island, where the uppermost clasts occur within the semi–pelite below the Doogart Schist and Limestone.

Thus in conclusion, the lithostratigraphy of the Tawneyshane Pelitic Member, including the absence of granitoid clasts, is within the lithological variation exhibited by the Port Askaig Tillite elsewhere in Ireland. The similarity of the lithology of the Tawneyshane Pelitic Member and its position within the stratigraphic succession is consistent with identification of this unit as the Port Askaig Tillite.

LECKEE QUARTZITIC FORMATION : In both the Ox Mountains and S. Donegal, the tillite passes up with increasing proportion and thickness of quartzite interbeds into a massive, impure quartzite unit, the Leckee Quartzite (Ox Mountains). The post–tectonic thickness of this unit in the Ox Mountains of 2300m suggests that this was originally a very thick quartzite unit. It is correlated with the Gaugin Quartzite (S. Donegal) on the basis of similarity in lithostratigraphy. This is one of the thickest and most persistent stratigraphic units in the British Isles, and is recognised in Scotland (Islay Quartzite, Islay and Jura; Schichallion Quartzite, Schichallion), and Ireland (Slieve Tooey Quartzite, Glencolumbkille and NW Donegal). Correlation of the Leckee Quartzitic Formation with the two other major quartzites in the British Dalradian can confidently be excluded as both of these are lithologically distinct from the Leckee Quartzitic Formation and have very different stratigraphic associations. The Killeter Quartzite (southern Donegal) of the Crinan Subgroup, the stratigraphic equivalent of the Lough Crana Quartzite (Inishowen and Fanad) lies stratigraphically above dark pelites of the Termon Pelite (the Stragill and Linsfort Black Schist at Inishowen and Fanad) and is overlain by the limestone and semi–pelite of the Aghyaran Formation (southern Donegal) and the Culdaff limestone (Inishowen and Fanad), which is a very different association to that observed in the Ox Mountains. The other possible correlation is equally difficult to support. The Ards Quartzite (northwest Donegal), the stratigraphic equivalent of the Appin Quartzite (Kintyre–Arran–Easdale–Appin) is underlain by the Ards Black Schists and overlain by the distinctive calcareous lithologies of the Sessiagh–Clonmass Group and the Falcarragh Pelite, a sequence which is very different form that observed in the Ox Mountains. Futhermore the Ar ds Quartzite itself is considerably coarser grained than the Leckee Quartzite and contains common pebble beds that are absent from the Ox Mountains.

LECKEE TRANSITION MEMBER : A transitional unit, which comprises a rapidly alternating sequence of calc-silicate schist with associated amphibolite, pelite, semi-pelite and dark psammite, can be recognised in both the Ox Mountains, (Leckee Transition Member) and S. Donegal (Reelan Formation).

UMMOON FORMATION : In S. Donegal the relatively large area of exposure reveals that large scale facies variations occur above the Gaugin Quartzite, and its upper transition, the Reelan Formation. No reliable stratigraphic marker horizons are present in this part of the succession; it is therefore difficult to confidently make a formation by formation correlation with the Ox Mountains. With increasing pelite the Reelan Formation passes upwards into the Croaghubbrid pelite comprising dark graphitic pelites, with thin ($<$ 1m) psammite bands and occasional lenses of dolomitic limestone. With an increasing proportion of psammite bands the Croaghubbrid Pelite passes up via a transition to the more massive psammites of the Boultypatrick Grit, that contains abundant blue quartz grains (maximum diameter 5mm) and localised occurrences of volcanogenic clasts (Alsop, 1987), which is the oldest evidence for volcanic activity in the Dalradian of northwest Ireland. The occurance of these volcanic clasts close to the Reelan Formation and underlying Gaugin Quartzite is consistent with the correlation of the Croaghubbrid Pelite and Boultypatrick Grit with the lower part of the Ummoon Formation of the Ox Mountains, which contains a thin metavolcanic unit, the Newantrim Member. Volcanics also occur at a similar stratigraphic level above the Islay Quartzite at Schichallion, at the top of the Killiecrankie Schist (Islay Subgroup), and above the Lakes Marble (Easdale Subgroup) in Connemara. They are absent from this stratigraphic level elsewhere in the British Isles.

In the Ox Mountains, the metasediments of the Ummoon Formation that occur above the volcanic member are similar to those below it, although the psammite bands are rather more frequent and slightly thicker. In S. Donegal the Boultypatrick Grit is succeeded by the Lough Eske Psammite and its lateral equivalent the Termon Pelite. Rapid facies variation occurrs at this level in south Donegal (Pitcher, Shackleton and Wood, 1971). The Lough Eske Psammite is characterised by thick beds of homogenous feldspathic psammite 1–1.5m thick, which contain less blue quartz than the Boultypatrick Grit. Limestone and calc-pelite lenses up to 1m thick and 30m in length are also recorded (Alsop, 1987). The grain size decreases towards the top of the formation and towards its lateral facies equivalent, the Termon Pelite, which is an extremely thick sequence (2km in S. Donegal) of thinly bedded muscovite rich pelites (some calcareous), psammites, greenbeds and quartzite. Alsop (1987) discovered greenbeds containing chlorite, epidote, albite, quartz, calcite and opaques within the formation and noted that the psammitic component of both the Termon Pelite and the Lough Eske Psammite has a green colouration throughout central and southern Donegal. The stratigraphically overlying Killeter Quartzite and Aghyaran Formation also show considerable lateral variation in thickness and lithology. When large scale facies changes such as this occur, lithostratigraphic correlation between widely separated areas can be extremely misleading, as similarity in lithostratigraphy does not necessarily imply chronostratigraphic equivalence.

LISMORAN FORMATION : In the Ox Mountains, the Ummoon Formation passes via a sedimentary transition into the Lismoran formation, that comprises an Upper and a Lower unit of feldspathic psammite, semi-pelite and pelite separated by a thick (200M maximum), metavolcanic unit - the Callow Member.

Volcanic units which are the only potential marker horizon in this part of the sequence, occur principally at two stratigraphic levels in the Dalradian of the British Isles. Extensive basic volcanic units occur in the Southern Highland Group, and at the top of the Argyll Group (Tayvallich Sub–group) of the Scottish Dalradian and at an equivalent stratigraphic level in Ireland (Shannaghy Green Beds, south Donegal; Greencastle and Cloghan Green Beds, Inishowen and Fanad). Less extensive volcanic units occur lower down in the succession (Farragon Beds, Easdale Sub–group, Blair Atholl; Sron Bheag Schists, Easdale Sub–group, Schichallion) in Scotland; and above the Lakes Marble (Easdale Sub–group, Connemara) in Ireland. Alsop (1987) discovered volcanogenic clasts in the Boultypatrick Grits, and Greenbeds within the psammitic lithologies of the Lough Eske Psammite and Termon Pelite of central and southern Donegal (Easdale Sub–group).

Two alternative lithostratigraphic correlatives are therefore possible for the Lismoran Formation and Callow Member. Occurrences of volcanic units at both stratigraphic levels in Ireland are within dominantly psammitic units that bear close lithostratigraphic similarity to the Lismoran Formation of the Ox Mountains. The older occurrance lies within the dominantly feldspathic psammite of the Lough Eske Psammite / Termon Pelite, while the younger occurrance lies between the feldspathic grits and psammites of the Mullyfa Grits and the finer grained semi–pelite and pelite of the Croaghgarrow Formation. Therefore neither correlation can be excluded on the basis of lithostratigraphic data.

It follows that if the correlation with the Shanaghy Greenbeds is correct, then the very thick Lough Eske Psammite / Termon Pelite of southern Donegal must be represented by a greatly reduced thickness of sediment in the Ox Mountains (the Ummoon Formation). Futhermore, the absence from the Ox Mountains of chronostratigraphic equivalents of the Aghyaran and Killeter Quartzite Formations requires considerable differences in sedimentation between this area and southern Donegal.

While significant facies variations have been demonstrated to occur in southern Donegal (Pitcher, Shackleton and Wood, 1971), the present author considers that the extreme facies variation that correlation of the Callow Member and the Shanaghy Green Beds requires, is unlikely. The simpler correlation with the Green beds of the Lough Eske Psammite / Termon Pelite which places all of the Ox Mountains Succession excluding the Cloonygowan Formation in the Argyll Group is therefore preferred.

3.1.2 Correlation of the Cloonygowan Formation with other areas of low- grade turbiditic metasediments

Lack of suitable marker horizons makes correlation of the Cloonygowan Formation with the other greenschist facies rocks of turbiditic character, that are exposed in a series of disconnected outcrops on the south side of Clew Bay and on Clare Island, difficult. This has led to a number of differing interpretations in the literature.

Phillips (1973) drew attention to the similarity of the Clew Bay Group of Clare Island and the south side of Clew Bay with the Highland Border Complex (HBC), that lies immediately to the to the south of the Dalradian rocks in Scotland. He did not, however, distinguish between these rocks and the Westport Grit and Cloonygowan Formations, which were previously regarded as belonging to the Southern Highland Group (SHG), Currall (1963), Taylor (1968). Long and Max (1977) correlated the Westport Grit Formation with the Cloongowan Formation and on the basis of similarity of lithostratigraphy assigned them to the SHG, which in Scotland is characterized by a thick sequence of mudstones, sandstones and rare conglomerates of submarine fan or turbidite facies. Max (1988), also regarded the Cloonygowan Formation to be equivalent to the green greywackes and turbidites of the Westport Grits but as being distinct from the Clew Bay Group which he defined as comprising the Ballytoohy Formation of Clare Island, and the Killdangan Formation of the south side of Clew Bay, and considered to be part of the HBC.

A detailed study of the rocks of Clew Bay and Clare Island are beyond the scope of the present work, however a brief reconnaissance of the Westport Grit Formation supports the correlation by Max (1989) of these rocks with the Cloonygowan Formation. The metasediments of the Cloonygowan Formation and the Westport Grits are both characterized by grey–green greywackes and quartzose turbidite grits, but lack the range of lithologies, particularly the dark mudstones, sedimentary breccias, and wide range of clasts in conglomerates of the Killadangan Formation. The rocks of the Cloonygowan Formation and Westport Grits also appear to share a common deformation history (chapter 4).

A fauna has not yet been recovered from the Ox Mountains, it is therefore not possible to establish their age or stratigraphic identity with confidence. The correlation of the Cloongowan Formation with the turbidites of the SHG is therefore based on similarity of lithostratigraphy alone.

3.1.3 Correlation of the Ox Mountains sequence with the Moine, (Phillips et al. 1975).

Phillips et al. (1975) consider that all the rocks of the Ox Mountains inlier, except for the small areas of low grade rocks accepted by all workers to be Upper Dalradian, are of Moinian age (Johnstone et al, 1969). They did not however attempt to make a lithostratigraphic correlation between the Ox Mountains and the Moinian of Scotland. This correlation is rejected by the present author on lithostratigraphic, structural and metamorphic grounds.

The lithostratigraphy of the central and southwest Ox Mountains is very distinct from that of the Moines, which is characterized by metamorphosed arenaceous and argillaceous sedimentary rocks which contain a very minor proportion of calcareous or dolomitic beds. The main rock types are thick homogeneous quartzo–feldspathic psammitic and pelitic schists, with intermediate semi–pelitic schist. In contrast, the metasediments of the central and southwest Ox Mountains are considerably more variable and include limestone and metavolcanic units. Acceptance of the whole of the Ox Mountains inlier as Moine therefore would imply a very different depositional environment from that of the Scottish Moines.

Phillips et al. (1975) cite structural evidence in support of their claim that the entire Ox Mountains inlier (except for the Cloonygowan Formation) is Pre-Caledonian. Their correlation of the first fabric developed in the Cloonygowan Formation rocks Sc1 (Sc2, present chronology) with D4 cataclasis in the higher grade rocks of the Ox Mountains sequence is however rejected on geometrical grounds. This is discussed fully in section 4.10. The present work has shown that that the deformation history in the Upper Dalradian rocks of the Cloonygowan Formation is considerably more complex than was previously recognised, and that the rocks of the Ox Mountains Succession and the Cloonygowan Formation share a common D1-D4 polyphase deformation history.

Phillips et al. (1975) suggest that the rocks of the central and southwest Ox Mountains are the products of retrogression of the high pressure granulites exposed in the northeast, or at least were deformed and metamorphosed under granulite facies conditions during their postulated first deformation event (D1). The metamorphic history of the rocks of the central and southwestern Ox Mountains is discussed in chapter 4, which shows that the present work is in agreement with the conclusions of Yardley et al. (1979). These workers concluded that the central and southwestern Ox Mountains have suffered a single progressive Barrovian metamorphism from the garnet zone to the Kyanite zone, and did not accept the evidence for retrogession from the granulite facies cited by Phillips et al. The correlation of the metasediments of the Ox Mountains Succession with the Moine is therefore not accepted by the present author.

3.2 SEDIMENTARY ENVIRONMENT AND SEDIMENT PROVENANCE

The intense deformation experienced by the metasediments of the Ox Mountains makes interpretation of the sedimentary environment, reconstruction of the original basin geometry and establishing the areas of provenance extremely difficult. Only a very limited amount of empirical data is available from the Ox Mountains themselves, most information has been gained therefore by extapolating information obtained from S. Donegal (Alsop, 1987).

The lithological characteristics of the Ox Mountains metasediments are consistent with the shelf environment proposed for the sediments of S. Donegal, whereas the coarse grained greywackes of the Cloonygowan Formation are characteristic of a proximal turbidite sequence. Neither palaeocurrent structures nor systematic variation in lithotype have been observed in either unit, it has therefore not been possible to determine the direction of sediment supply. By analogy with S. Donegal it appears likely that the metasediments of the Ox Mountains Succession were supplied from as high grade metamorphic source, which is possibly part of the N Mayo-NE Ox Mountains-Lough Derg basement block.

The source of the sediments of the Cloonygowan Formation is equally poorly constrained. Several possible sources have however been suggested for the Cloonygowan Formation and its corrolatives, the Westport Grit Formation and, on a larger scale, the Southern Highland Group (SHG) as a whole. The presence of granitic pebbles, quartz, microcline and sodic plagioclase led Sutton and Watson (1955), and Borradaile (1973) to suggest a granitic or gneissic terrane as a source. This has been thought of as partly charnokitic, because of the universal presence of blue quartz grains, Wilson (1882), Sutton and Watson (1955). Phillips (1973) proposed a southern source for these turbidites, as he considered that the thick clastic succession could not have been derived from the shelf area to the north west, on which the Durness Limestone accumulated. If this were the case, post depositional sinistral movement on the Highland Boundary Fault System would have removed the source, which would now be situated along strike to the north east.

Anderton (1985) preferred to regard the sediment supply as coming from the northwest and stated that due to poor palaeontological dating, it was difficult to assess the significance of the Durness carbonate blanket as the youngest SHG turbidites could be entirely older than the oldest part of the Durness Limestone. In addition, he pointed out that the Nd isotope data of O' Nions et al. (1983) made no distinction between the SHG source terrane and the rest of the Dalradain. These conclusions were also supported by Kneller (1987).

If the sediments have been derived from the northwest, the areas of high grade granulite facies rocks of north Mayo, the northeast Ox Mountains and Lough Derg must be considered to be the most probable source. Their derivation from the amphibolite facies of the central and southwest Ox Mountains, suggested by Phillips (1975), is considered to be unlikely, due to the absence of material from which the blue quartz that is ubiquitous in the SHG could be derived.

3.3 CONCLUSIONS

Identification of the Boulder Bed within the Ox Mountains is of critical importance. In addition to resolving the question of the age of the metasediments of the central and southwest Ox Mountains, it has a number of other important implications.

It enables the metasediments to be correlated with the stratigraphic successions established in Dalradian rocks in the rest of the British Isles. In particular, the close correlation between the Ox Mountains metasediments and those of the Dalradian succession in S. Donegal, whose way up and age is known, strongly supports the stratigraphic succession established in the Ox Mountains where sedimentary way up evidence is largely absent. In the following chapters, the metasediments of the Ox Mountains sequence will therefore be referred to as the Ox Mountains Succession.

The correlation with the upper part of the S Donegal succession, which is stratigraphically continuous, i.e. not dismembered by major tectonic slides, is inconsistent with the model of terrane accretion proposed by Winchester, Long and Max (1988), which necessitates a major terrane boundary to separate the Ummoon and Lismoran Formations in the Ox Mountains.

CHAPTER 4

DEFORMATION CHRONOLOGY

Previous structural studies of the central and southwest Ox Mountains have lead to a number of contrasting interpretations of the structure of parts of the inlier and a number of widely differing deformation chronologies. Many of these differences result from the limited amount of data available to previous workers who, individually, studied only small parts of the inlier, rather than a difference in observations. No previous attempt has been made to integrate data from the entire inlier to constrain its structural and kinematic development. Figure 4.1 summarizes the deformation chronologies presented by previous workers and the deformation chronology that has resulted from the present study of the whole inlier. This information, which is integrated with metamorphic studies, provides the data on which the tectono-thermal model of chapter 5 is based.

A detailed study of the deformation of the igneous complex is beyond the scope of the present project. However in order to understand fully the structural development of the inlier, it is necessary to appreciate the basic structural history of the three igneous bodies that intrude the metasediments. The structural history of the igneous complex is therefore reviewed briefly in section 4.9. While most of these data result from reconnaissance studies made by the writer in 1986 and subsequent detailed examination of the contact relationships between the igneous complex and the country rocks in 1987-1989, the present author's understanding of the structure of the interior of the igneous complex has benefitted from discussion with K. McCaffrey. All information derived from these discussions is acknowledged in the text that follows.

SUMMARY OF DEFORMATION CHRONOLOGIES

FIG 4.1

CURRALL (1963) SW Ox Mts Succn	
D1	F1 Recumbent folds
D2	F2 overturned folds, SW plunge, strain slip cleavage
D3	F3 upright folds SE plunge
D4	brittle thrusting from NW monoclinal congugate folding

TAYLOR (1969) Central Ox Mts Succn.	
D1	Major Nappe Folds. Axial pl. traces NNE-SSW. LTS.
D2	Upright tight-close folds, NE-SW trending. GWS. L. Talt antiform (F2)
	OX MTS GD LTA
D3	Congugate folds, A NNE-SSW trending B E-W " Attymass Synform
D4	Kink bands assoc. open folds, NW-SE trending.

PHILLIPS ET AL (1977) Central Ox Mts Succn		Cloonygowan Fm	
?	Pre S1 qtz segretations		
D1	Minor NE-SW F1 folds, included S1 fabric.		
D2	isoclinal NE-SW folds, S2 penetrative fabric, LTS.		
	OX MTS GD LTA LEA		
D3 ②	Close NE-SW folds, LT Antiform, S3 crenulation fabric, Ist fabric in OX MTS GD.		
D4	S4 cataclasis	$D_{c}1$	Penetrative fabric, Fc1 folds. contact with Ox Mts succn.
D5	NE-SW Upright Knockalongy antiform	$D_{c}2$	Fc2 NE-SW synform
		$D_{c}3$	Fc3 monoclinal warping

LONG AND MAX (1977) SW Ox Mts Succn		Cloonygowan Fm	
D1	S1 included fabric, isoclinal folds. upward facing L Anaffrin antiform.		
D2	F2 tight-isoclinal folds, S2 ax pl fabric.		
⑤	OX MTS GD Late D2		
④	First fabric in OX MTS GD, coplanar with regional D2.		
D3	2nd fabric in GD. cataclastic contact between Ox Mts succn and Cloonygowan Fm.	$D_{c}1$	penetrative schistosity S1 cataclastic at L Rusheen. no major folds identified.
D4	Minor flexuring, kink bands, fracture cleavage.	$D_{c}2$	Minor folding, S2 local crenulations.
		$D_{c}3$	Minor flexuring, kink bands, fracture cleavages.
③	LEA LTA		

ANDREWS (1978) Central Ox Mts Succn			
D1	Qtz segregations		
D2	F2 folds highly attenuated profiles, S2 ax pl cleavage.		
	AMPHIBOLITE INTRUSION		
D3	F3 folds S3 intense fabric mostly composite S1-S3. stretching lineation parallel to fold axes.		
D4	F4 open-tight folds. Knockalongy antiform, no penetrative fabric.		
	OX MTS GD		
D5	Rewoking in NE zones, open-tight congugate angular folds, and spaced fabric. L E S.	$D_{c}1$	
D6	Congugate angular folds in NE block	$D_{c}2$	

HUTTON AND DEWEY (1986) C Ox Mts, Cloonygowan Fm	
D1	
D2	
D3	tectonic contact with Ox Mts Succn. Intense spaced fabric. Sinistral shear bands developed close to contact with Ox Mts Succn.
	OX MTS GD

PRESENT STUDY C & SW Ox Mts succn,		Cloonygowan Fm	
D1	Quartz segregations preserved in hinges of rare F2 folds.	D1	S1 grain shape alignment fabric
D2	Rare isoclinal F2 folds, preserved in F3 hinges strong gently inclined penetrative fabric, Vergence indeterminate.	D2	Rare isoclinal F2 folds, strong penetrative fabric, at small angle to bedding. quartz pebbles strongly flattened.
PROGRESSIVE D3	Tight-isoclinal F3 folds verge consistently towards the GD. S3 spaced fabric largely transposes S2. strong gently plunging stretching lineation. S3 intensifies towards LTS, GWS, CSZ and NOXS.	PROGRESSIVE D3	Open-isoclinal gently plunging F3 folds F3 folds tighten towards Intense S3 spaced fabric, sinistral shear bands deueloped close to the tectonic contact with the Ox Mts Succn, (CSZ).
	OX MTS GD		
	Sc3 Shear bands developed in OX GD and Ox Mts Succn.		
D4	congugate folds with angular profiles, axes sub-parallel to L3 stretching lineation.	D4	Congugate folds with angular profiles, axes sub-parallel to L3 stretching lineation.
	LEA LTA		
D5	Sc5 sinistral shear bands developed in LTA LEA. late dextral shears. Kink bands.	D5	Dextral kink bands with steeply plunging axes.

① State that the two fabrics in the GD cannot be correlated with fabrics in the country rocks.

② States in footnote that this fabric in the GD may be S4-Sc1

③ Later with O'Connor 1984 describe intrusion of LEA, ~~LTA~~ as post D4

④ State correlation is equivocal

⑤ Later regarded as syn-post F3, Long and Yardley (1979)

⑥ Andrews 1984 describes LEA as intruded into a DEXTRAL shear zone

Abbreviations

GD	granodiorite
LEA	L. Easky Adamellite
LTA	L. Talt Adamellite
LTS	L. Talt Slide
GWS	Glennawoo Slide
NOXS	North Ox Mountains Slide
CSZ	Callow shear zone

In this chapter the methodology of structural and metamorphic analysis employed is justified (section 4.1) and the terminology outlined (section 4.2) prior to a detailed examination of the structural and metamorphic data from each of 6 areas, (sections 4.3-4.8.), which is summarized in figures4.2-3.The structure of the components of the igneous complex is reviewed in section 4.9 and finally the evidence for the correlation of the deformation and metamorphic sequences of the Ox Mountains Succession and the Cloonygowan Formation is presented in section 4.10.

4.1 METHODOLOGY OF STRUCTURAL ANALYSIS

A deformation chronology which can be recognised and correlated throughout the area has been established by examining overprinting relationships in the field, supported by microscopic examination of thin sections. This sequence of deformation has been dated relative to the emplacement of the Ox Mountains Granodiorite, Lough Talt and Lough Easky Adamellites and forms the basis for the structural and kinematic analysis that follows.

The first deformational event (D1) is associated with a specific set of tectonic structures such as folds (F1) and cleavage (S1). The next phase (D2) produces F2 and S2 etc. Bedding is not a tectonic structure and is designated Ss, except where it has been extensively modified by deformation and is termed lithological banding. The intersection of cleavage associated with a particular deformation phase (Dn) with bedding (Sn/Ss), or with a preceeding tectonic fabric (Sn/Sn-1), is parallel to fold axes (Fn) of that age. Both structures are referred to as β lineations (β). Throughout the area mineral elongation lineations (stretching lineations), (Xn), associated with a particular event are sub-parallel to the fold axes (Fn) produced by the same deformation. Therefore they are also plotted as β lineations. Extensional crenulation cleavages or shear bands related to a

REGIONAL DEFORMATION CORRELATION (with peak regional metamorphic assemblages)

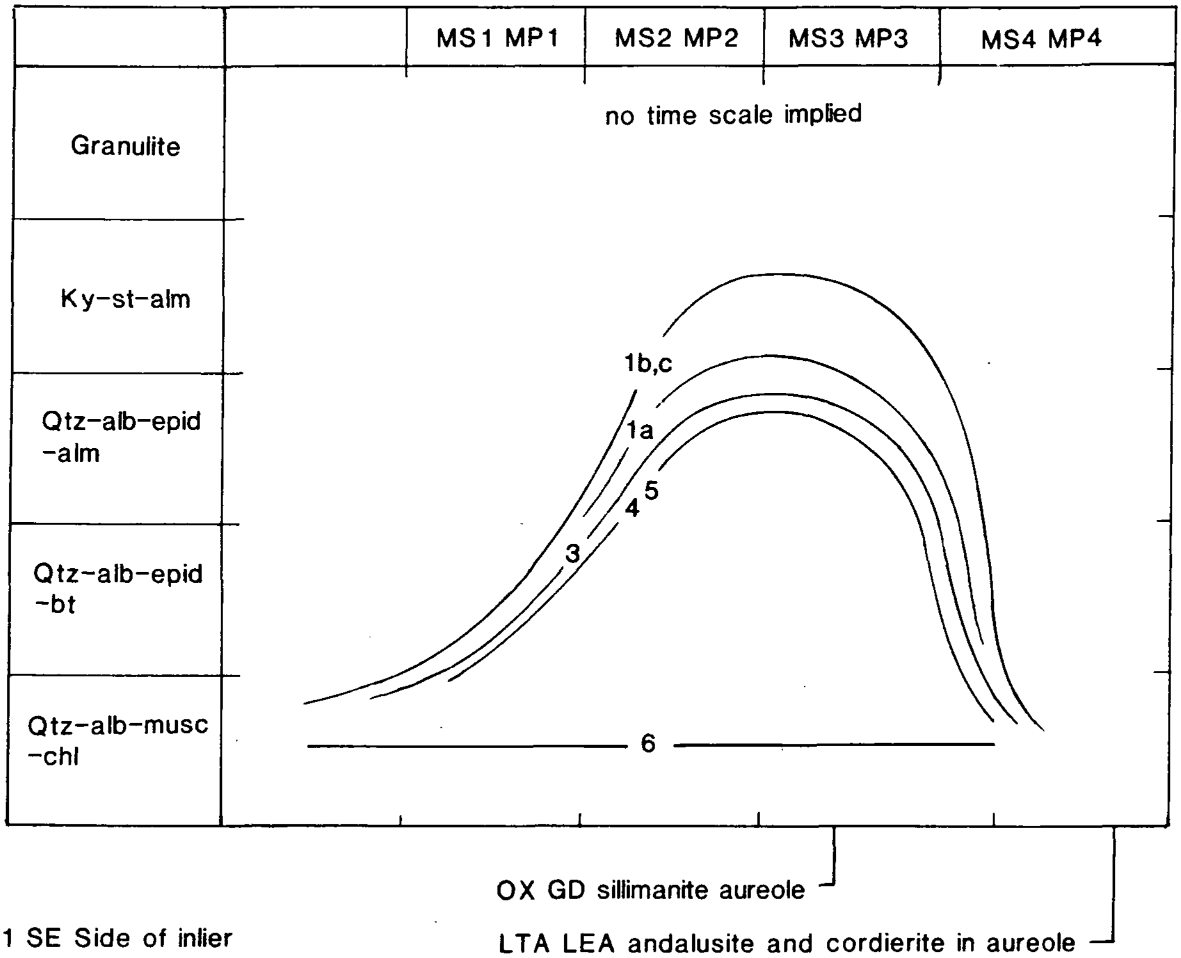
FIG 4.2

	SE SIDE OF INLIER (AREA 1A)	(AREA 1B)	(AREA 1C)	LOUGH TALT (AREA 2)	NW SIDE OF INLIER (AREA 3A)	(AREA 3B)	SHANVOLLEY (AREA 4)	TULLYCOMMONS (AREA 5)	CLOONYGOWAN (AREA 6A)	ARDVARNEY (AREA 6B)	WESTPORT (AREA 6C)
D1	S1 penetrative fabric rarely preserved in F2 fold hinges. Elsewhere S1 is transposed.	Si penetrative fabric rarely observed in F2 fold hinges .	S1 not observed	S1 penetrative fabric, observed rarely in F2 fold hinges.	S1 penetrative fabric observed rarely in F2 fold hinges.	S1 penetrative fabric rarely preserved in F2 fold hinges	S1 not observed	S1 not observed	S1 grain shape alignment fabric.	S1 grain shape alignment fabric.	S1 not observed
D2	S2 penetrative fabric preserved in F3 fold hinges elsewhere transposed F2 rare, upright folds, preserved in the hinges of co-axial F3 folds. MP2 st gt, bt, musc, feld, qtz.	S2 penetrative fabric preserved in F3 fold hinges, elsewhere transposed. F2 rare isoclinal folds, preserved in F3 fold hinges, co-axial with F3 MP2 ky, st, gt, feld, bt, musc.	S2 penetrative fabric preserved in F3 fold hinges, elsewhere transposed. MP2 st, gt, musc, bt, qtz.	S2 penetrative fabric preserved in F3 fold hinges, elsewhere transposed. F2 rare folds, vertical NE-SW trending axial planes, axes plunge vertically. MP2 st, gt, musc, bt, qtz.	S2 penetrative fabric preserved in F3 fold hinges, elsewhere transposed. F2 rare isoclinal folds preserved in f3 fold hinges, co-axial with F3, MP2 gt, st, musc, bt, feld, qtz.	S2 penetrative fabric preserved in F3 fold hinges, elsewhere S2 is transposed MP2 st, gt, musc, bt, feld.	S2 penetrative fabric MP2 gt, feld, bt, musc, qtz. F2 folds not observed	S2 penetrative fabric MP2 gt, bt, musc, feld	S2 penetrative fabric. F2 NE-SW trending, recumbent isoclinal folds, face SE. MS2-MP2 musc & qtz original tectonic contact with the Ox Mts Succn. ?	S2 penetrative fabric. F2 recumbent isoclinal folds MS2-MP2 musc, chl, qtz. Tectonic contact with the Ox Mts Succn. S2C sinistral shear bands .	S2 penetrative fabric, MP2 musc, chl, qtz. Contact with the Ox Mts Succn. is not observed.
D3	S3 intense penetrative fabric, NE-SW strike, sub-vertical orientation, intensifies towards the tectonic slides. F3 upright, tight-isoclinal folds, vergence is consistently towards the granodiorite, axes plunge gently, parallel to X3. LTS, GWS, CSZ. SC3 sinistral shear bands .	S3 intense penetrative fabric, NE-SW trending, dips moderately NW. F3 upright, tight-isoclinal folds, vergence consistently SE, axes plunge gently NE & SW, THSZ. S3C sinistral shear bands developed throughout the area.	S3 intense penetrative fabric, sub-vertical in SE, shallower dips in the NE of the area. F3 upright, tight-isoclinal folds, axes plunge gently parallel to X3 Variable vergence. Fabric intensifies towards the NOXS. SC3 shear bands rare.	S3 intense penetrative fabric, NE-SW trending, dips moderately-steeply NW. F3 upright, tight-isoclinal folds, verge towards major antiformal hinge (The Lough Talt Antiform) SC3 sinistral shear bands occasionally observed.	S3 intense penetrative fabric, NE-SW trending, dips moderately to NW. F3 upright, tight-isoclinal folds, vergence consistently SE, axes plunge gently NE & SW. SC3 sinistral shear bands developed throughout the area.	S3 intense NE-SW trending fabric, shallowly inclined. SC3 sinistral shear bands	S3 intense penetrative fabric, moderately-steeply inclined. F3 upright, tight-isocinal folds, vergence is towards the LTS & GWS. SC3 shear bands observed throughout the area.	F3 folds, rare, upright, axes plunge gently N or S No pentrative fabric	S3 pressure solution cleavage NE-SW trending, subvertical orientiation. F3 upright tight-isoclinal folds, axes plunge gently NE. Tectonic contact with the Ox Mts Succn. SC3 sinistral shear bands.	No penetrative fabric	S3 pressure solution cleavage NE-SW trending, subvertical orientation. F3 upright, tight-isoclinal folds with axes plunging gently NE & SW.
D4	F4 open-close angular folds, axes parallel to X3, axial planes shallowly inclined .	As area 1a	As area 1a	As area 1a	As area 1a	As area 1a	As area 1a	As area 1a	As area 1a	F4 open-close angular folds, axes parallel to X2. Axial planes shallowly inclined.	As area 1a
D5							Fractured monoclinal folds.		Dextral kink bands	Fractured monoclinal folds.	

Abbreviations LTS - Lough Talt Slide GWS - Glennawoo Slide CSZ - Callow Shear Zone NOXS - North Ox Mountains Slide ODZ - Oblique Drop Zone OBZ THSZ - Tawnaneillen High Strain Zone

ky - kyanite gt - garnet feld - feldspar musc - muscovite bt - biotite qtz - quartz chl - chlorite

FIG 4.3



1 SE Side of inlier
3 NW Side of inlier
4 Shanvolley
5 Tullycommons
6 Cloonygowan Fm

GENERALIZED METAMORPHIC HISTORIES OF THE INLIER

Modified after Long & Max (1977)

deformation event (Dn), and produced by intense strain along a pre-existing foliation, are termed (Cn), (after Cisaillement, or shear) Berthé et al. (1979).

The deformation chronology established in this way has been integrated with the metamorphic history by thin section studies of the relationships of porphyroblasts to their internal (Si) and external (Se) fabrics. Syn-tectonic mineral growth associated with deformation phase (Dn), is termed (MSn), while mineral growth that postdates Dn is designated (MPn) etc.

The concept of D numbers provides a useful framework for the systematic description of structures. In the Ox Mountains the present author has used the concept to divide the deformation into a series of events that reflect distinct kinematics. It should however be emphasised that they are arbitrarily defined markers on a deformation and metamorphic path that that is considered to be largely progressive.

The term progressive deformation was first proposed by Flinn (1962, pp. 387-388) who suggested that rocks undergoing progressive deformation will pass through a continuous series of shape changes until deformation ceases. In the Ox Mountains this term is used by the present author to refer to the D3 deformation so as to imply–

1. A close relationship between the various D3 structures in terms of orientation, sense of movement, style and prevailing metamorphic conditions.

2. A relatively constant orientation of the regional stress field.

3. The various structures developing as a relatively continuous sequence of events within a geologically limited period of time.

In areas of high strain, this progressive deformation has resulted in the production of a composite D1-D3 fabric which is defined as a 'planar surface or group of surfaces whose components share similar morpological features and orientations' (Tobisch and Patterson, 1988). In the Ox Mountains, the composite D1–D3 fabric has been developed by a combination of the following mechanisms: rotation of the early D1 and D2 foliations into the new D3 orientation, rotation of early folds parallel to X3, by transposition of S1 and S2 by S3, recrystallization and formation of neoblasts parallel to S3.

4.2 MICROSTRUCTURE NOMENCLATURE

Structures associated with the high strain zones and tectonic slides in the Ox Mountains have consistently been referred to by previous workers as 'cataclastic'. They interpreted the irregular or seriate boundaries of the coarse, deformed quartz grains and islands of large grains and porphyroclasts in a matrix of finer grained material, as suggesting brittle deformation. Thus it has been inferred that these structures developed late in the tectonic history of the inlier, at lower metamorphic grade than the ductile deformation that produced folding and fabric development. The purpose of this section is not to review the extensive literature on deformation mechanisms (see Knipe, 1988) but to examine briefly the features that characterise deformation in shear zones under various deformation mechanisms, and then to examine in more detail the microstructure of the inlier so as to evaluate the PT conditions and the genetic relationship of these structures to other fabrics within the inlier.

The term mylonite was first used by Lapworth (1885) to describe the fine grained rocks in the Moine Thrust Zone in Eriboll. The term originally implied grain size reduction by brittle processes and this has remained the universal view until recently. It has been noted frequently that the milled rock had a recrystallized texture which was thought to result from post-tectonic grain growth. Extensive grain growth was also invoked to explain the presence of large, apparently undeformed, porphyroblasts. More recent microstructural studies of the Moine Mylonites (Bell and Etheridge, 1973) have indicated that recrystallization was syntectonic. This has led to a clear division between cataclasites and mylonites (White, 1976). Grain refinement in cataclasites is accomplished by fracture and subsequent sliding and rotation, whereas in the case of mylonites, grain refinement is by recrystallization or new mineral growth with subsequent deformation being by ductile processes. In the case of polymineralic rocks the behaviour of the matrix alone is considered; hence, although feldspar megacrysts in the high strain zones in the igneous lithologies in the Ox Mountains are commonly fractured, these rocks still reflect mylonitic deformation.

White (1980) states that the development of mylonitic microstructure with increasing shear strain proceeds with an increase in the proportion of recrystallized grains relative to porphyroclasts. In a monomineralic mylonite, the porphyroclasts that are most unfavourably orientated for slip resist recrystallization and form globular porphyroclasts which eventually recrystallize, whereas those that are perfectly orientated tend to form tabular or ribbon grains parallel to the foliation. In polymineralic rocks such as the feldspathic psammites and semi-pelites of the Ox Mountains, minerals mechanically stronger than the matrix tend to remain as porphyroclasts, that are often extended and fractured parallel to the extension direction.

The recrystallization and new mineral growth that produces most mylonites, occurs preferentially around the edges of the original grains, as these edges have external boundary conditions imposed upon them. New grains therefore develop by strain induced boundary migration and sub-grain rotation. The preservation of small grains in mylonites such as the Callow Shear Zone and the LTS and GWS west of the Knockaskibole Fault, indicates that little growth has occurred after recrystallization.

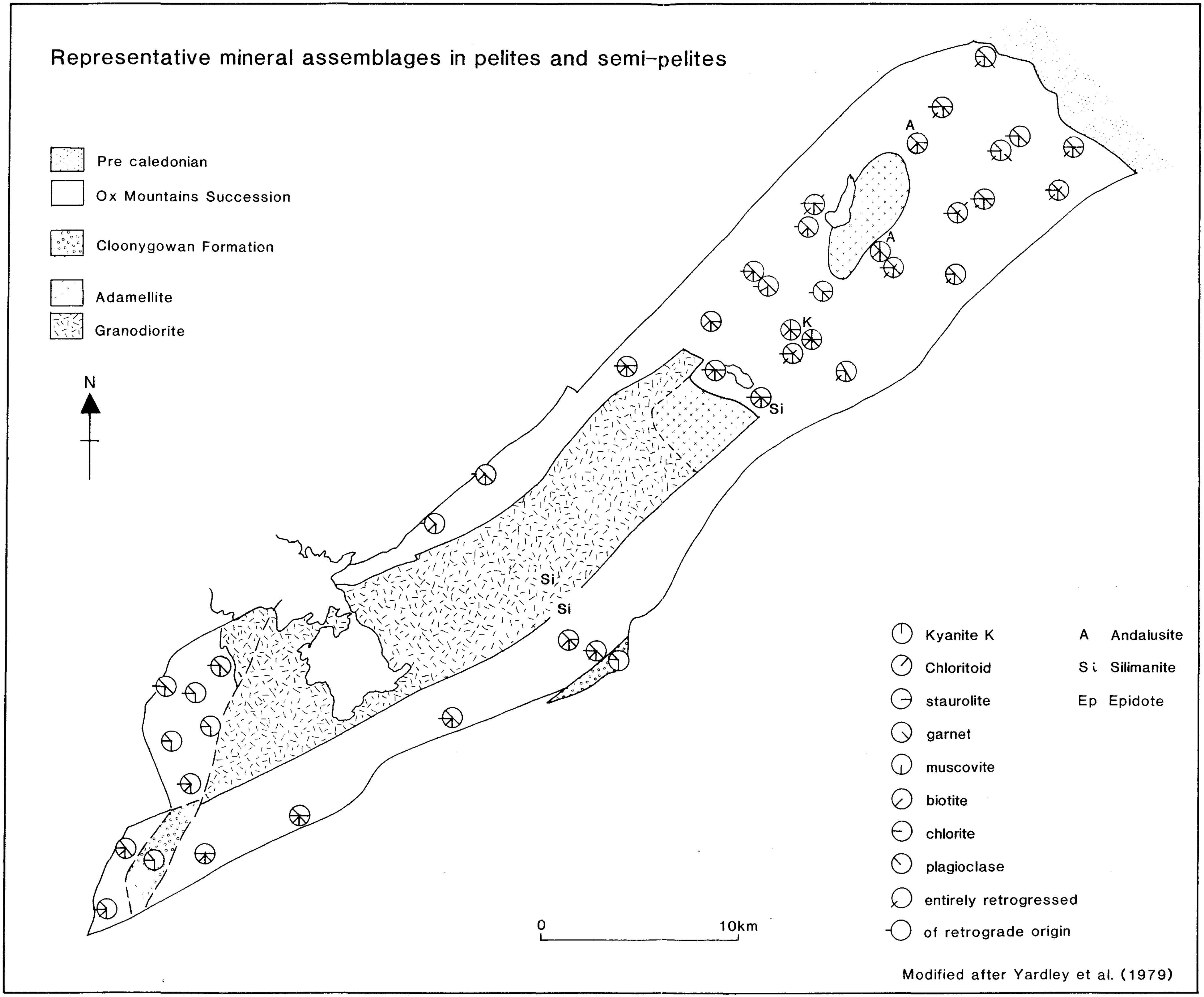
At high strains a late non penetrative foliation commonly develops at a low angle (< 35°) to the mylonitic foliation. The two fabrics intersect with a pronounced asymmetry and can therefore be used to determine the shear sense. These late fabrics occur widely in the Ox Mountains, where they are termed shear bands, and are assigned the nomenclature SCn, where S is the foliation that they deform and C is the shear band, named after 'Caisaillment' or shear. The significance of these features is that they developed during the same deformation as the mylonite but at a later stage. White et al. (1980) state that it is probable that the shear bands represent the final phase of ductile deformation in the shear zone as the temperature drops and uplift commences and they are therefore often associated with retrogressive metamorphism. This observation is consistent with data from the Ox Mountains, where shear bands and the associated fabric which they deform show a very close spatial correlation with retrograde metamorphism, and the growth of chlorite. Garnets are chloritized and are frequently broken and extended parallel to the mylonitic foliation. Quartz has a strong preferred dimensional orientation while feldspar has remained largely undeformed. This retrograde metamorphism (Figure 4.4) is distinct from regional retrograde metamorphism produced during the uplift of an orogenic zone, where rocks remain essentially undeformed and lack extensive recrystallization and fabric development. The presence of water during retrograde metamorphism is considered by Beach (1980) to play an important part in shear zone initiation. In the Ox Mountains the shear bands commonly show the following features:

1. Shear bands are commonly inclined at 30-40° to the mylonitic foliation.

2. The apparent sense of displacement on the cleavage consistently indicates a component of extension along S.

3. The shear bands are discontinuous and tend to curve into and anastamose with S. The length of the individual shear band is normally considerably greater than the wavelength of the crenulation.

FIG 4.4



Representative mineral assemblages in pelites and semi-pelites

Modified after Yardley et al. (1979)

4. Where observed the intersection of the shear band with S makes a high angle (>70°) to the stretching lineation associated with the mylonitic foliation. The intersection is usually curvilinear.

5. In thin section the shear bands are distinguishable as very narrow (approximately 50μ) zones of intense microstructural modification. The grain size of the quartz is reduced from an average size of ($125\times250\mu$) to ($25\times25\mu$). Micas also show a grain size of ($25\times250\mu$) compared with ($125\times325\mu$) outside the zone. Micas at the margins of the zones are bent into the shear band. There is no noticable change in bulk compostion within the shear bands. The lack of chemical differentiation is used by Platt and Vissers (1980) to infer that volume was approximately conserved, and therefore that the displacements are real and that the C planes have caused a significant extension along the earlier foliation.

The final texture of strained rocks is largely a function of the interplay between strain rate and recovery. The recovery processes involve the release of strain energy accumulated in the crystal lattices. This is commonly accomplished by syn-tectonic recrystallization often with a reduction in grain size or by late post-tectonic annealing resulting in equi-dimensional grains. Two critical factors in the preservation of microstructures are the rate of stress drop at the end of the deformation event and the temperature history that post-dates the deformation. The rapid removal of the environmental conditions which induce microstructural changes (ie rapid stress or temperature drops) will also enhance the preservation of microstructures of high deformation. This is clearly illustrated by the microstructure of the GWS and LTS. West of the Knockaskibole fault these structures are characterized by extensive sub-grain growth and ribboning of quartz grains, while east of the fault the same structures have experienced less rapid fall in temperature and more recovery has taken place. This gives the erroneous impression that deformation in the latter area was less intense than in the former.

4.3 AREA 1: SE SIDE OF THE INLIER

This is the largest area of the inlier and corresponds to the lithostratigraphically defined area 1 (Figure 2.2). The area is of great importance as it exposes the boundaries between all the stratigraphic units in the Ox Mountains. It includes 7 of the 8 major tectonic discontinuities in the inlier and contains much of the critical evidence for establishing a deformation chronology, and timing of emplacement of the plutons. These data therefore provide the basis for the comparison of the structural and metamorphic history of the rest of the inlier.

In this section the critical overprinting relations of structures on the southeast side of the inlier are examined and the geometry of SS, D1 and D2 structures reviewed for the entire area. The area is then subdivided into three smaller sub-areas to illustrate the systematic change in geometry of the D3 structures.

Sub-area 1a extends from the Knockaskible fault in the southwast (section 4.6.5), to the townland of Carrowmore (G 3550 0540) 5km northeast of Callow, where a large poorly exposed area separates it from the central sub-area, (1b). The sheeted contact with the granodiorite (Section 4.3.6) forms the northwestern limit of the ground, while Carboniferous rocks onlap to the southeast. Sub-area 1b extends from Carrowreagh (G 3850 1200) southeast of Glennawoo to the Mad river (G 4850 1800) in the northeast. In the southwest part of the area the Ox Mountains Granodiorite forms its northwest margin. Further to the northeast, area 1b is in contact with area 2 at Lough Talt. Northeast of Lough Talt, a large area of blanket peat isolates the area from rocks of the northwest side of the inlier (area 3). Sub-area 1c is bounded to the northeast by the tectonic contact with the northeast Ox Mountains (section 4.3.5.6) and to the northwest by a large area of blanket peat, which separates it from the northwest side of the inlier (area 3). Its southern contact is defined by the boundary between the Ummoon Formation and the Lower Lismoran Formation (The Glennawoo Slide, section 4.3.5.2). The area is overstepped by Carboniferous to the southeast.

Exposure in this the largest area of the inlier is best on the large, steep, north-facing slopes of Glennmore and Knockachree which provide an almost continuous vertical northwest-southeast section through the northern part of the central Ox Mountains. Additional exposure is provided by the stream sections that flow east into the river Moy.

4.3.1 Ss Lithogical banding

Throughout areas 1a and 1b lithological banding lies parallel to the penetrative S2 fabric (Plate 4.1) which in areas of high D3 strain has itself been largely transposed by S3. This results in a simple outcrop pattern in the areas a and b, where intense transcurrent deformation has produced a steeply inclined S3 fabric and all of the stratigraphic contacts are now sub-vertical. The S3 foliation and lithological boundaries have a shallower dip in area 1c. At Knockachree a gently dipping stratigraphic contact between the Ummoon Formation and the Lower Lismoran Formation shows little evidence of tectonic modification.

4.3.2 D1

In common with many areas that have experienced a protracted history of pervasive and often intense deformation, structures associated with D1 are extremely rarely preserved as they have been transposed by later deformation. The earliest fabric observed in the field is a quartz, muscovite segregation. This is preserved in the hinges of F2 folds, which themselves are quite rare and only clearly identifiable where refolded by F3. The best examples of this relationship are preserved in the semi-pelite and pelite of the Ummoon Formation at Ummoon (G 2715 0007) (Plates 4.2–4.3). In thin section this fabric is a very fine grained (<0.05mm) MS1 fabric consisting of quartz and muscovite which is preserved as inclusions in MP2 albite porphyroblasts. No lineations have been observed associated with the S1 fabric, therefore no vergence information has been obtained. S1 is parallel to Ss but neither associated intersection or stretching lineations are

Plate 4.1 Lithological banding in the Lismoran Formation, parallel to the S2 penetrative fabric, folded by upright, northwest verging F3 folds. (Ummoon G 2875 0036).

Plate 4.2 Small scale isoclinal F2 fold folding S1 quartz segregations, refolded by tight, upright F3 folds. Ummoon Formation, Ummoon (G 2715 0007).

observed and therefore the original orientation and relationship of D1 structures with respect to lithological banding are unknown.

4.3.3 D2

The second deformation is represented by an intense penetrative fabric consisting of an MS2 segregation of phyllosilicates, predominantly muscovite and biotite (now largely retrogressed to chlorite) and quartzo-feldspathic layers approximately 0.5mm thick. It appears therefore that metamorphic grade lay within the greenschist facies prior to MP2. The MS2 fabric is positively identified in the hinges of F3 folds. Elsewhere where S3 is intense S2 is transposed and forms a composite S1-S3 fabric. Development of the S2 fabric was followed by the growth of porphyroblasts of plagioclase and garnet throughout the area. These commonly enclose fine-grained, straight inclusion trails of muscovite, quartz and opaques. Large kyanite porphyroblasts up to 2cm long, containing inclusion trails of quartz and opaques are developed in pelites of the Ummoon Formation at Zion Hill (G 4240 1740). The external fabric (Se=S3) forms augens around these porphyroblasts confirming that they are of MP2 age. The peak of regional metamorphism is therefore constrained to be of MP2 age. Transposition by the later D3 has generally reduced the angle between S2 and Ss to the point that it cannot be resolved. Hence no vergence information of S2 to bedding can be obtained. If the effects of F3 folding are removed it appears that S2 was originally shallowly inclined throughout the area.

F2 folds are rare, and are generally observed as small scale isoclines (maximum wavelength 50cm), refolded by co-axial F3 folds with which they form type 3 interference patterns (Plates 4.2–4.3). A large number of larger scale F2 folds were invoked by Andrews et al. (1978) from the Cabbragh-Ladies Brae area to account for the stratigraphic relationships that they observed. However they did not cite any structural evidence in support of this. Mapping by the present author

Plate 4.3 Small scale isoclinal F2 fold folding S1 quartz segregations, refolded by tight, upright F3 folds. Ummoon Formation, Ummoon (G 2715 0007).


10p

has not found supporting structural data and has resulted in the re-identification of stratigraphic units in this area (see chapter 2).

As D2 structures have been transposed it has not been possible to verify the existence of large scale F2 folds using vergence data. The existence of F2 structures such as the Lough Anaffrin Antiform (Long and Max, 1977) has therefore been inferred from stratigraphic evidence alone (outlined in section 2.1.2.).

Apart from the Lough Anaffrin Antiform the general absence of stratigraphic repetitions or relationships that cannot be explained simply in terms of D3 or later structures, suggests that other major F2 folds are probably absent. D2 in the inlier appears to have been largely a fabric forming event, and did not produce a complex series of nappe structures analogous to those observed in the stratigraphically equivalent area of S. Donegal (Alsop, 1987).

4.3.4 D3

AREA 1a : In this area, D3 is represented by a strong spaced fabric and common tight-isoclinal folds with wavelengths commonly between 1cm and 1m and corresponding amplitudes of between 3cm and 3m. S3 is a steeply inclined fabric with a mean orientation of 055/85SE (Figure 4.5), consisting of a preferred dimensional and crystallographic orientation of quartz and phyllosilicates. The fabric is a composite fabric in which components of S2 can be identified in microlithons between S3 foliation planes. In high strain zones, S2 is almost totally transposed and makes a small angle to S3, with the vergence of S3 on S2 consistently towards the northwest. S3 is modified by shear-bands which have a consistently sinistral sense of displacement over the entire area, (Plate 4.4).

Tight-isoclinal F3 folds with sub-vertical axial planes and gently plunging axes have a mean plunge orientation 28/057, parallel to the very strongly developed stretching lineation, defined by a strong dimensional and crystallographic orientation of quartz (Plate 4.5), and verge consistently to the northwest. No vergence

Plate 4.4 Sinistral shear bands, Lismoran Formation, Callow Post Office (G 3232 0257).

Plate 4.5 Outcrop of Lismoran Formation at Lismoran (G 3232 0257) illustrating the strongly developed, gently northeast plunging stretching lineation, on sub–vertical S3 foliation surfaces. (Outcrop is approximately 5m high).

Figure 4.5 Stereographic projection of structural data : Area 1 (S.E. Side of the inlier)

Area 1a :

(a) S3 : mean orientation of foliation plane 055/85, (pole, 05/325), n=384

(b) β3 : mean orientation of lineation 28/057, n=284

(c) S4 : mean orientation of axial plane 306/21 (pole, 69/216), n=104

(d) β4 : mean orientation of lineation 31/052, n=52

Area 1b

(a) S3 : mean orientation of foliation plane 087/66 (pole, 24/307), n=373

(b) β3 : mean orientation of lineation 05/228, n=159

(c) S4 : mean orientation of axial plane 328/38 (pole, 52/238), n=341

(d) β4 : mean orientation of lineation 30/026, n=229

Area 1c

(a) S3 : mean orientation of foliation plane 190/21 (pole, 69/100), n=135

(b) β3 : mean orientation of lineation 11/329, n=106

(c) S4 : mean orientation of axial plane 283/47 (pole 43/193), n=15

(d) β4 : mean orientation of lineation 19/285, n=12

Percentage of data points enclosed by contours.

☐ >2%
🟨 >5%
🟩 >10%
🟧 >15%
🟦 >20%
⬛ >25%

FIG 4.5

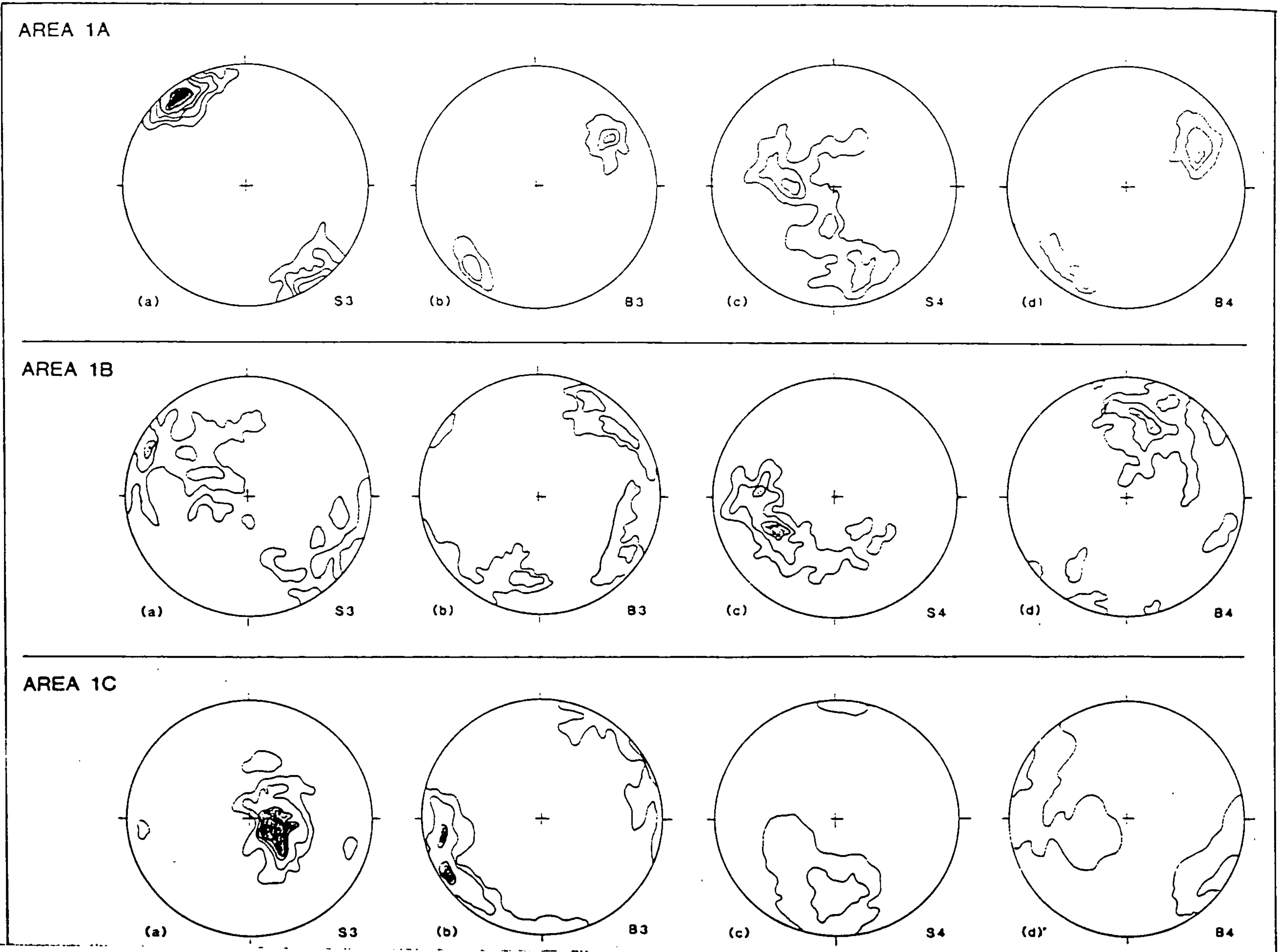
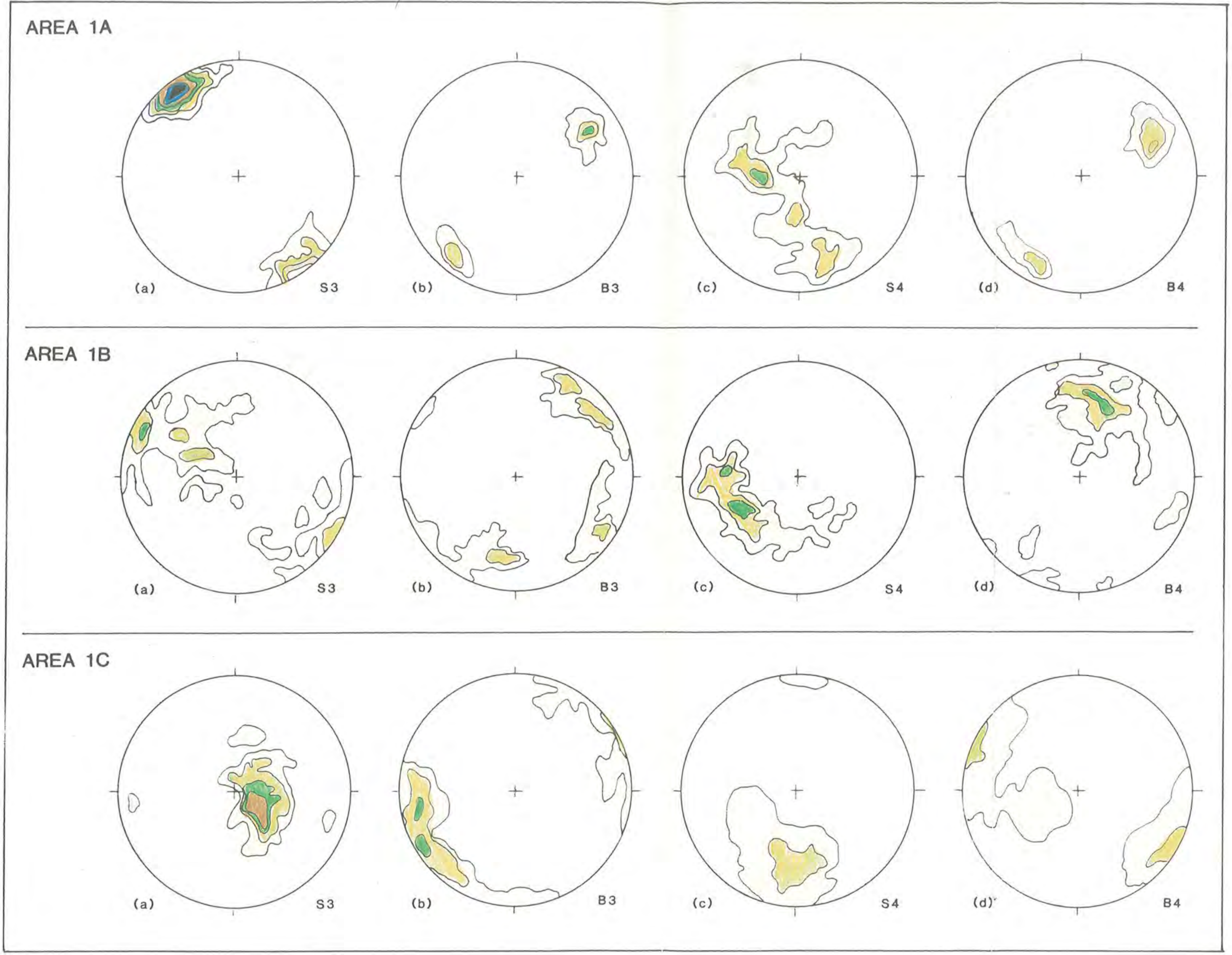

AREA 1A
(a)
S3
(b)
B3
(c)
S4
(d)
B4
AREA 1B
(a)
S3
(b)
B3
(c)
S4
(d)
B4
AREA 1C
(a)
S3
(b)
B3
(c)
S4
(d)'
B4

FIG 4.5

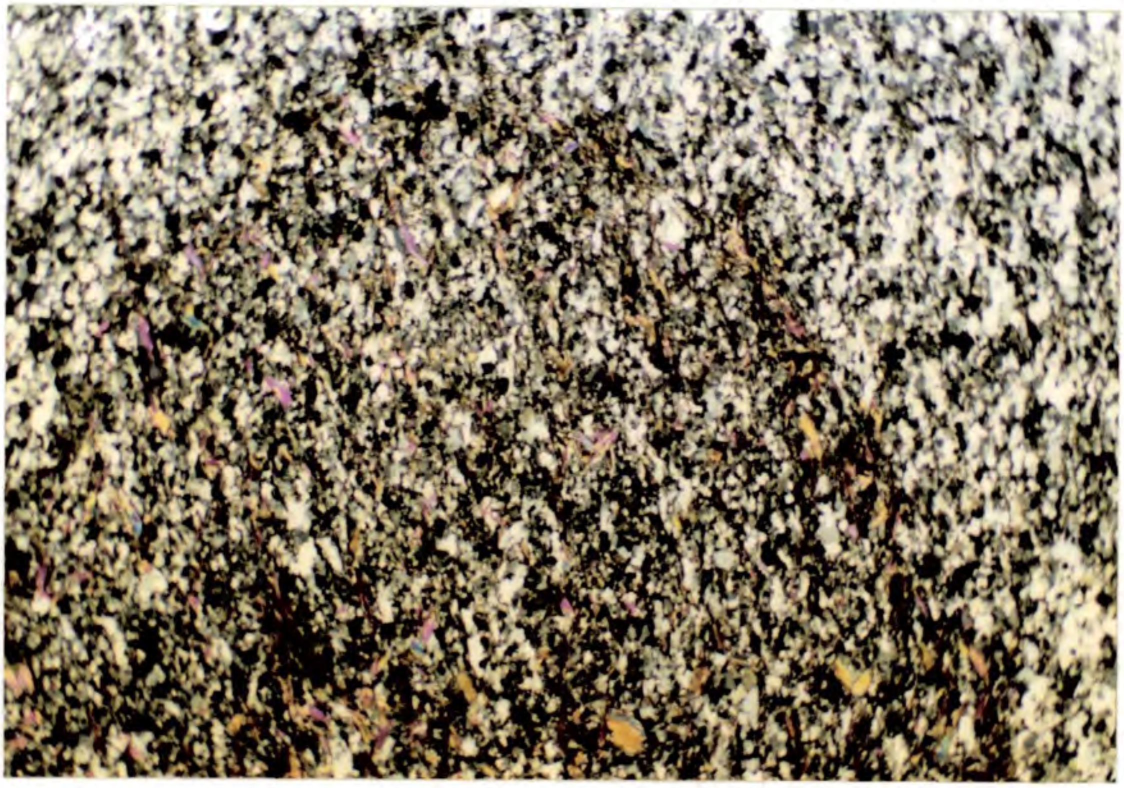

AREA 1A
(a)
S3
(b)
B3
(c)
S4
(d)
B4
AREA 1B
(a)
S3
(b)
B3
(c)
S4
(d)
B4
AREA 1C
(a)
S3
(b)
B3
(c)
S4
(d)
B4

data have been observed to support the presence of parasitic F3 folds such as the Kilmore Antiform and Callow Antiform described by Taylor (1968). In F3 hinges S3 is a preferred dimensional orientation of the (001) planes of muscovite and biotite which crenulate earlier micas (Plate 4.6). MS3–MP3 muscovite and biotite have grown over S3 foliation planes, and quartz has been removed by pressure solution. In the limbs of F3 folds MS3 micas form a composite S1–S3 schistosity. MP3 garnet growth often forms an outer inclusion free rim to MP2 garnets, while large euhedral MP3 staurolite randomly overgrows MS3 micas. Chloritoid growth is restricted to area 4c where it occurs as randomly orientated porphyroblasts that overgrow MS3 micas.

A qualitative assessment of the D3 strain is provided by the spacing of S3, the presence or absence of shear-bands and the degree of tightness of F3 folds. This indicates that strain is inhomogeneously distributed along the southeast side of the inlier (Figure 4.6) and that three high strain zones or tectonic slides are present. Plates 4.7–4.9 illustrate the progressive tightening of F3 folds and intensification of S3 approaching the Glennawoo Slide. The detailed structure of the tectonic slides is discussed in greater detail in section 4.3.5.

AREA 1b : Structures related to D3 in this area show similar geometric and metamorphic relations to those in area 1a to the southwest, However, the strike of the S3 fabric swings through approximately 60° from a mean orientation of 057/83NE at Glennawoo (G 4020 1286) to a mean orientation of 105/52S at Killcummin (G4900 0000), (Figure 4.5). The present author considers that this is likely to be related to the effects of the Dalradian metasediments of the Ox Mountains Succession being driven over a gently south dipping basement-cover interface. This is reflected by the D3 stretching lineation which most commonly dips gently to the southwest or west. F3 vergence is more variable than in area 4a, suggesting the presence of mesoscopic parasitic folds. Southeast verging isoclines have been identified but no major folds have been observed. Most of the vergence

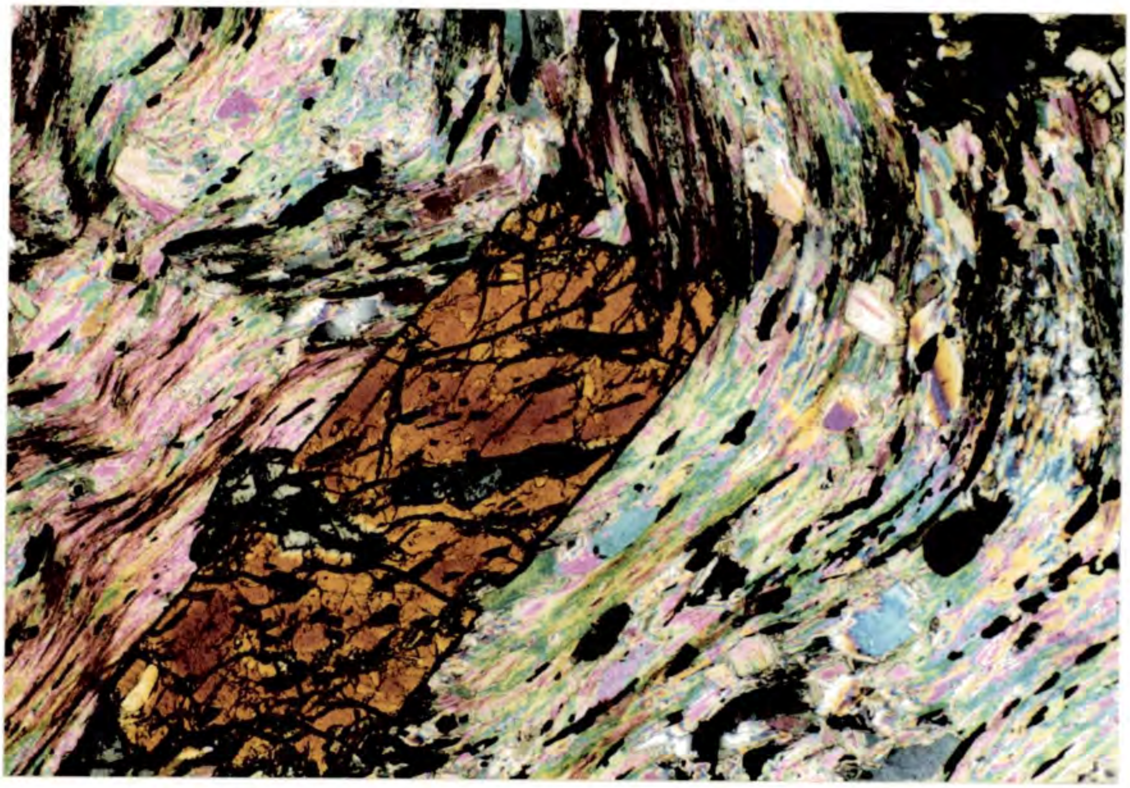
Plate 4.6 Photomicrograph of F3 fold hinge, illustrating S2 orientated parallel to lithological banding, crenulated by the steeply inclined S3 fabric, developed axial planar to upright F3 folds. Ummoon (G 2715 0007).



Field of view 14mm

Plate 4.7 Northwest verging , upright F3 folds, Lismoran Formation, 200m southeast of the Glennawoo Slide, (G 2880 0041).

Plate 4.8 Northwest verging , upright F3 folds, Lismoran Formation, 110m southeast of the Glennawoo Slide, (G 2875 0048). Note increase in tightness and amplitude from Plate 4.7.



Field of view 3.5mm

data suggest that the area is on the southeast limb of an antiform whose axis lies to the northwest.

AREA 1c : The northern part of the area was studied by Andrews et al. (1978), who mapped a sinuous boundary between the Ummoon and Lismoran Formations, and described a complex structural model involving stratigraphic repetition by a large number of F2, F3 and F4 folds. Remapping by the present author (Chapter 2) has eliminated the requirement for many of the structures inferred by Andrews et al., and found that there is no structural evidence supporting their model.

The S3 foliation, which is a composite S1-S3 fabric, is recognised throughout the area. It is defined by a preferred dimensional and crystallographic orientation of MS3 muscovite, and green or red-brown biotite layers that are separated by quartzo-feldspathic layers that have subseqently recrystallized and exhibit no preferred dimensional or crystallographic orientation. In pelites S3 is commonly overgrown by MP3 staurolite, garnet and plagioclase (Plate 4.10). This indicates that metamorphism remained in the amphibolite facies field for a considerable time following the MP2 peak of regional metamorphism.

The S3 fabric describes a gentle antiformal structure plunging gently southwest, whose axis is located near Glennmore (G 5250 2550). This controls the outcrop pattern of the area. Neither minor structures or mineral growth features are associated with this large structure. The geometry of the D4 structures, which have variable trends (Figure 4.5), indicates that they have formed as a result of vertical shortening. These structures are clearly superimposed upon the large antiformal structure. Therefore they either formed at a late stage in the development of the antiform or postdate it entirely. The timing of development of the large scale antiform is therefore closely constrained to be either syn- or post–D3 but pre–D4. The geometry of the structure and these constraints on the timing of its development are consistent with it having been produced by sinistral transpression, which drove the metasediments of the area over the basement-cover interface that dips to the south or southeast (see chapter 5).

In general, in the area to the northwest of the fold axis the S3 foliation dips gently (<25°) northwest. The axes of tight F3 folds here plunge gently southwest and lie parallel to a moderately developed stretching lineation. Vergence of minor F3 folds is inconsistent throughout this area, suggesting the proximity to a major northwest closing recumbent F3 fold. At Knockachree (G 5170 2930) psammite, semi-pelite, pelite and metavolcanics of the Ummoon Formation are exposed in the steeply inclined cliff section above Lough Achree. These units, which are repeated by a large number of mesoscopic, gently plunging F3 folds, lie structurally below outcrops of pale weathering psammite and semi-pelite of the Lismoran Formation. As discussed earlier there is no indication that the boundary between them is tectonic.

At Glennmore to the southeast of the major gentle fold axis the strike rotates from a north-south orientation immediately north of Glennmore (G 5250 2550) to a northeast-southwest orientation at Cabbragh. Moving to the southwest of Cabbragh S3 becomes progressively steeper, reaching an orientation of 115/56S at (G 5200 2156) (the river Moy), where the contact between the Ummoon and Lismoran Formation is exposed. Southeast of this contact, which describes an arcuate trend between the River Moy and Glennawoo the structure and stratigraphy clearly represent a continuation of the southeast side of the inlier to the southwest. Evidence of sinistral shear (shear-bands, rotated porphyroblasts and mica fish) is common, and F3 folds, whose axes lie parallel to the gently plunging stretching lineation, consistently verge to the northwest, as with areas 1a and 1b.

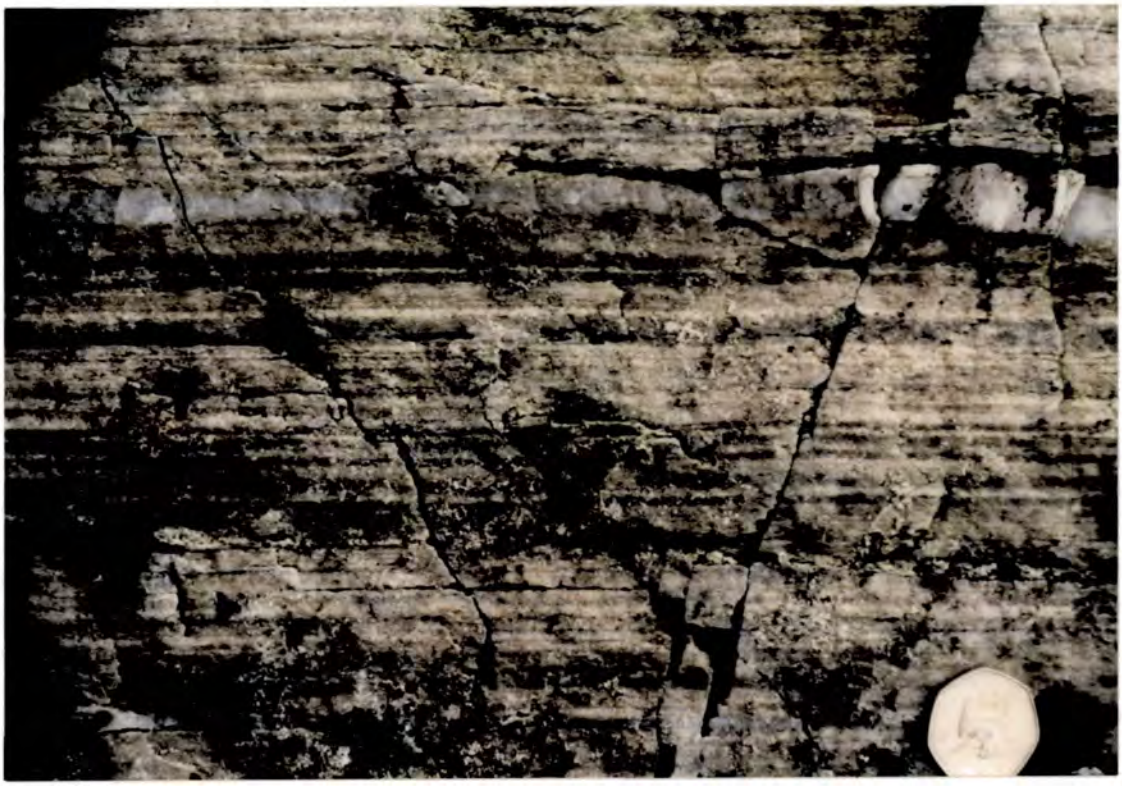
4.3.5 The tectonic slides and high strain zones

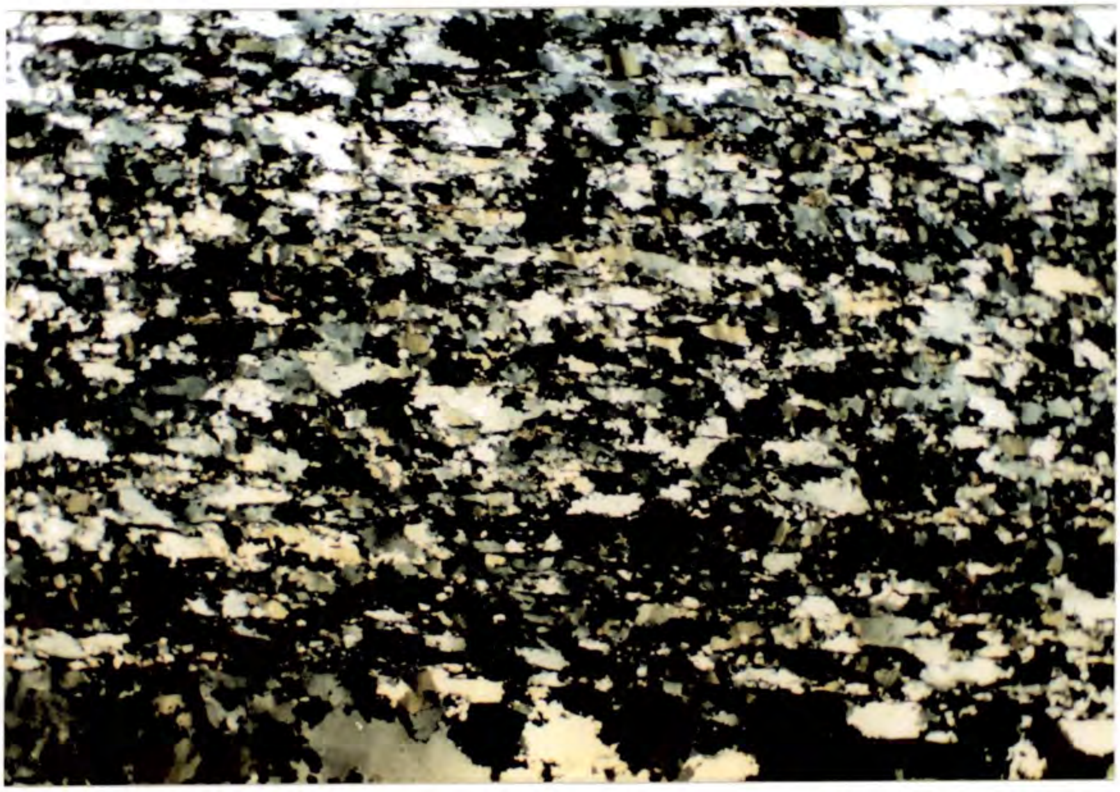
4.3.5.1 Lough Talt Slide : The Lough Talt Slide was named by Taylor (1968) who mapped it from sporadic exposures along the southeast side of the granodiorite between Ummoon (G 2800 0077) and Lough Talt (G 3975 1459). Taylor considered that the structure was contained within the Leckee Quartzitic formation for almost all of its exposed length, but that in the Ummoon area it cuts out the basal part of the Ummoon Formation. It is unclear from Figures 1 and 7 of Taylor (1968) what he regards the geometry of the structure to be at Lough Talt. The structure of this area is examined in detail in section 4.4. It appears however that he regards the slide to be folded around the Lough Talt Antiform, which is the major F3 (present chronology) antiform forming the core of the inlier , and that the slide here separates the Leckee Quartzitic Formation from the underlying Attymass Pelite Group. Phillips et al. (1975) stated that the slide separated rocks of different metamorphic grade, since they interpreted the diopside, microcline mineralogy of the Kilmore Limestone (Leckee Quartzitic Formation) as high amphibolite or granulite facies, and the Ummoon formation exposed to the southeast as low amphibolite facies. The conclusion of Long and Max (1977b) that this statement is invalid as neither diopside or microcline are diagnostic of the high amphibolite facies, (Turner and Verhoogen, 1960), is supported by the present work which has shown that no contrast in metamorphic grade can be demonstrated to occur across the slide. Thus, garnet, staurolite pelites occur within the Leckee Quartzitic Formation at Callow (G 3058 0393) northwest of the slide and also in the Ummoon Formation southeast of it near Lough Akista (G 3055 0299).

The slide is exposed at Ummoon (G 2840 0077) where it is identifiable as a 10m thick zone of intense D3 strain. This is reflected by the tightening of F3 folds and the development of an intense planar S3 fabric, accompanied by the total transposition of D2 structures (Plate 4.11). The latter effect makes it impossible

Plate 4.11 Mylonitic S3 foliation in the Leckee Quartzitic Formation, the Lough Talt Slide, Ummoon (G 2840 0076).

Plate 4.12 Thin section of sample from outcrop illustrated in plate 4.11, illustrating the strong preferred dimensional orientation of quartz and subgrain growth.





Field of view 14mm

to determine whether the structure had an earlier, possibly equally important, D2 tectonic history. At this locality the slide separates the Leckee Quartzitic Formation from the Ummoon Formation, and no evidence was observed that the structure cuts across stratigraphy at this locality. In thin section (Plate 4.12), the Leckee Quartzite shows a very strong preferred crystallographic orientation of grains which have intense sub-grain growth along their boundaries. The grains also have a preferred dimensional orientation, parallel to the contact with the Ummoon Formation, and x:y ratios of approximately 3:1. Minor sub-grain growth indicates that the fabric here has not totally recrystallized . The pelitic and semi-pelitic rocks of the Ummoon Formation immediately southeast of the slide are highly deformed. Shear bands are developed that consistently indicate a sinistral sense of displacement, along the X3 stretching lineation that here plunges gently to the southwest.

The structure is also exposed at Callow Bridge (G 3150 0333) where it has a similar expression. A large exposure gap of approximately 15km then separates this area from the next exposure at Lough Talt on the south side of Crummus (G 3975 1459), where it again separates the Leckee Quartzitic Formation from the Ummoon Formation. The amount of horizontal and vertical displacement on the slide cannot be determined, but the structure cuts out a minimum of several hundred metres of the Leckee Transition Member at Lough Talt, which suggests that the total displacement on the structure is likely to be large.

Blanket peat cover prevents tracing of the structure to the northeast. It is possible that it either dies out or merges with either the high strain zone that passes through the Lough Easky Adamellite (the oblique drop zone of chapter 5) or the northeasterly continuation of the Glennawoo Slide.

4.3.5.2 The Glennawoo Slide: The Glennawoo Slide was named after its type locality at Glennawoo (G 4020 1286) by Taylor (1969), who recognised that it was an intensification of the regional D2 fabric (D3 in the present chronology).

The slide separates the Ummoon and Lismoran Formations. Phillips et al (1975), however did not regard this boundary as tectonic and described it as a normal stratigraphic contact. Long and Max (1977), working in the southwest of the area mapped the slide and they regarded it as D3 in age (present D4) separating the Lismoran and Ummoon Formations at Shanvolley.

Andrews et al. (1978), traced the trajectory of the Lough Easky Slide which they considered to be the northeasterly continuation of the Glennawoo Slide, to the northeast where it was cut by the Lough Easky Adamellite. Andrews et al. described the slide as post D4 in age (current chronology) and suggested that the structure was an important late zone of sinistral Grampian reworking that postdated the intrusion of the granodiorite. These workers described the fabric of the Lough Easky Adamellite as resulting from approximately 2.5km of dextral displacement, but did not attempt to explain the inconsistency between this and the sinistral displacement on the Lough Easky Slide. It is now known that these workers misinterpreted the shear sense in the pluton by inferring shear sense from fractures in feldspar megacrysts that are antithetic to the overall shear sense.

Winchester et al. (1987, figure 1) show the slide forming a sinuous boundary between the upper and lower Lismoran Formations mapped by Andrews et al. (1978). However in reply to Jones and Leat (1988), Winchester et al. acknowledged that two additional possibilities existed for its trajectory northeast of Lough Talt. The first, that the slide dies out or diverged northwards from its interformational position to the southwest of the Lough Easky Adamellite, and lies entirely within the Ummoon Formation, was rejected as it was inconsistent with their hypothesis that the Glennawoo Slide was a major terrane boundary that everywhere seperated the Ummoon and Lismoran Formations. They therefore preferred the second option that the Glennawoo Slide can be traced out of the inlier approximately 1.5 km east of the Mad river, beneath the Carboniferous cover sequence with lack of exposure obscuring its actual course.

Mapping by the present author confirms that the slide is locally expressed as a zone of increased D3 strain (above the high background level of deformation observed on the southeast side of the inlier). In this section the location of the slide, its structural and metamorphic expression will be examined from the four areas of best exposure. The significance of the structure for the tectonics of the inlier and the conclusion of Winchester et al. (1988) that this structure represents a major terrane boundary will be reviewed critically in chapters 5 and 6.

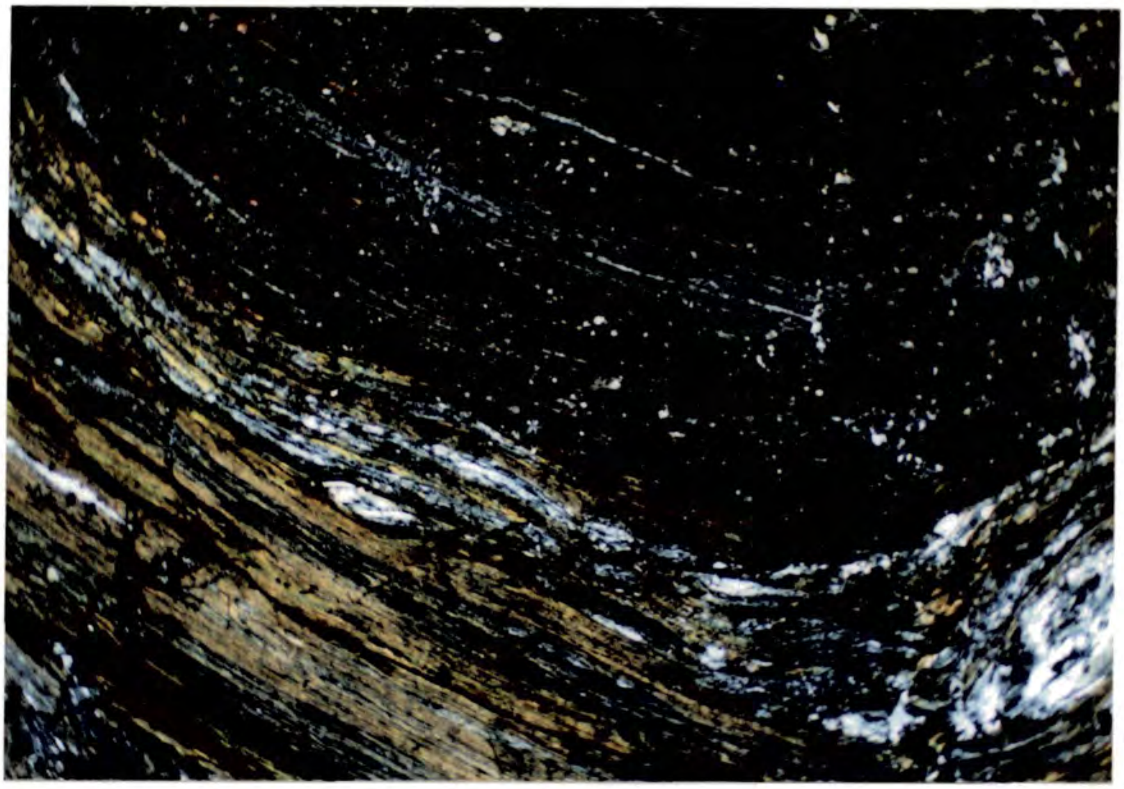
The type locality of the GWS is located 1500m SW of Glennawoo Bridge (G 4145 1369) where very highly deformed feldspathic psammites and semipelites of the Lismoran and Ummoon Formations are exposed in a series of small (maximum size 8 × 15m) outcrops that are isolated by large areas of no exposure. In hand specimen (Plate 4.13) the feldspathic psammite and semi-pelite are expremely fine grained and white in colour, due to the kaolinization of the feldspar. The steeply inclined mylonitic foliation that contains a strong gently plunging stretching lineation, is folded by conjugate F4 folds. In thin section (Plate 4.14) layers of quartz ribbons (less than 0.01mm wide), exhibiting axial ratios of >25:1 are separated by thinner layers of phyllosilicate (<0.01mm wide) composed of muscovite, chlorite, quartz subgrains and rare feldspar porphyroblasts.

It is unclear where the continuation of the Glennawoo Slide lies to the northeast of the type locality at Glennawoo, as fabrics of comparable intensity are not observed elsewhere in the inlier. The first option outlined by Winchester et al. (1988), that the course of the slide follows the sinuous boundary between the Ummoon Formation and the Lismoran Formation as mapped by Andrews et al. (1978) would mean that the slide is highly discordant with the orientation of the S3 fabric in the area. This is inconsistent with the fact that the slide is of D3 age (as shown at Glennawoo and elsewhere).

The second option, that the slide continues with a northerly trajectory towards Lough Easky is consistent with the observed structural data. Furthermore

Plate 4.13 Mylonitic S3 foliation in psammite, deformed by angular F4 folds, the Glennawoo Slide, Glennawoo (G 4020 1286).

Plate 4.14 Thin section from the same locality as Plate 4.13 illustrating the extreme axial ratios of quartz ribbons, separated by layers of ultra fine– grained phyllosilicates.



Field of view 14mm

reactivation of a structure with this orientation could account for the sinistral deformation fabric observed in the Lough Easky Adamellite.

The third option, that the slide follows the orientation of the S3 foliation, swinging to the east out of the inlier east of the Mad river is the simplest hypothesis. This hypothesis simply requires that the slide remains parallel to S3 throughout its exposed length and therefore also lies parallel to the revised boundary between the Ummoon and Lismoran Formations.

In summary both the second and third options are consistent with the observed structural data, and neither can be discounted. The present author considers that most of the displacement on the slide is likely to be taken up by deformation parallel to the regional orientation of S3. It is also possible that the slide may bifurcate and a small component of displacement may be taken up on a dislocation that passes through the Lough Easky Adamellite.

9km of very poorly (< 1%) exposed ground separates the type locality of Glennawoo from the well exposed Callow-Ummoon area, where a zone of high strain is located at a similiar stratigraphic level to that at Glennawoo. In outcrop the narrow zone of increased strain is marked by intense development of sinistral shear bands and a decrease in spacing of the S3 foliation, (Plate 4.15). The planar mylonitic fabric observed at Glennawoo is, however, not present. In thin section the dislocation is marked by the development of seriate boundaries to quartz grains and the growth of subgrains along the shearbands (Plate 4.16). Chlorite is concentrated along the shear planes, reflecting the retrogression of biotite in the slide zones. There is no evidence of an abrupt change in lithostratigraphy or metamorphic grade across the slide zone Jones and Leat (1988).

The boundary between the Ummoon and Lismoran Formations is exposed at Greenans (G 1705 0117). It is extremely difficult to locate the GWS as high strain fabrics and intense sinistral shear bands are present throughout the area.

Plate 4.15 Thin section of psammite of the Ummoon Formation, adjacent to the Glennawoo Slide illustrating tight F3 folds, Callow Loughs (G 3160 0333).

Plate 4.16 Thin section of same specimen as Plate 4.15 cut parallel to the stretching lineation to illustrate the development of shear bands that modify the S3 foliation. Sub grains of quartz are extensively developed. Chlorite is concentrated along the shear planes, reflecting the retrogression of biotite to chlorite in the slide zones.

Field of view 14mm

Field of view 14mm

Identification of the GWS is therefore made by noting the slight increase in intensity of deformation that coincides with the boundary between the Ummoon and Lismoran Formations arbitrarily defined by the incoming of bands of psammite >1m thick.

4.3.5.3 The Callow Shear Zone (CSZ): The 1000m wide zone of intense deformation comprising high strain fabrics and sinistral shear bands that occurs along the southeast side of the inlier, is termed the Callow Shear Zone (CSZ) by the present author. This separates the Upper Lismoran Formation of the Ox Mountains Succession from the Cloonygowan Formation. Within the zone the actual contact between these units is everywhere concealed beneath Carboniferous cover. Phillips et al. (1975) inferred this contact to be a major Dc1 (D5 present chronology) slide zone, characterised by cataclasis and retrograde metamorphism in the Upper Lismoran Formation. Hutton and Dewey (1986) also recognised that the boundary between the two units was a steep shear zone, but noted that the shear zone fabric is an intensification of S3 in both sets of rocks. They stated that the high grade unit – the Ox Mountains Succession, is brought up against the low grade Cloonygowan Formation by an intervening D3 sinistral shear zone with a gently plunging transport lineation.

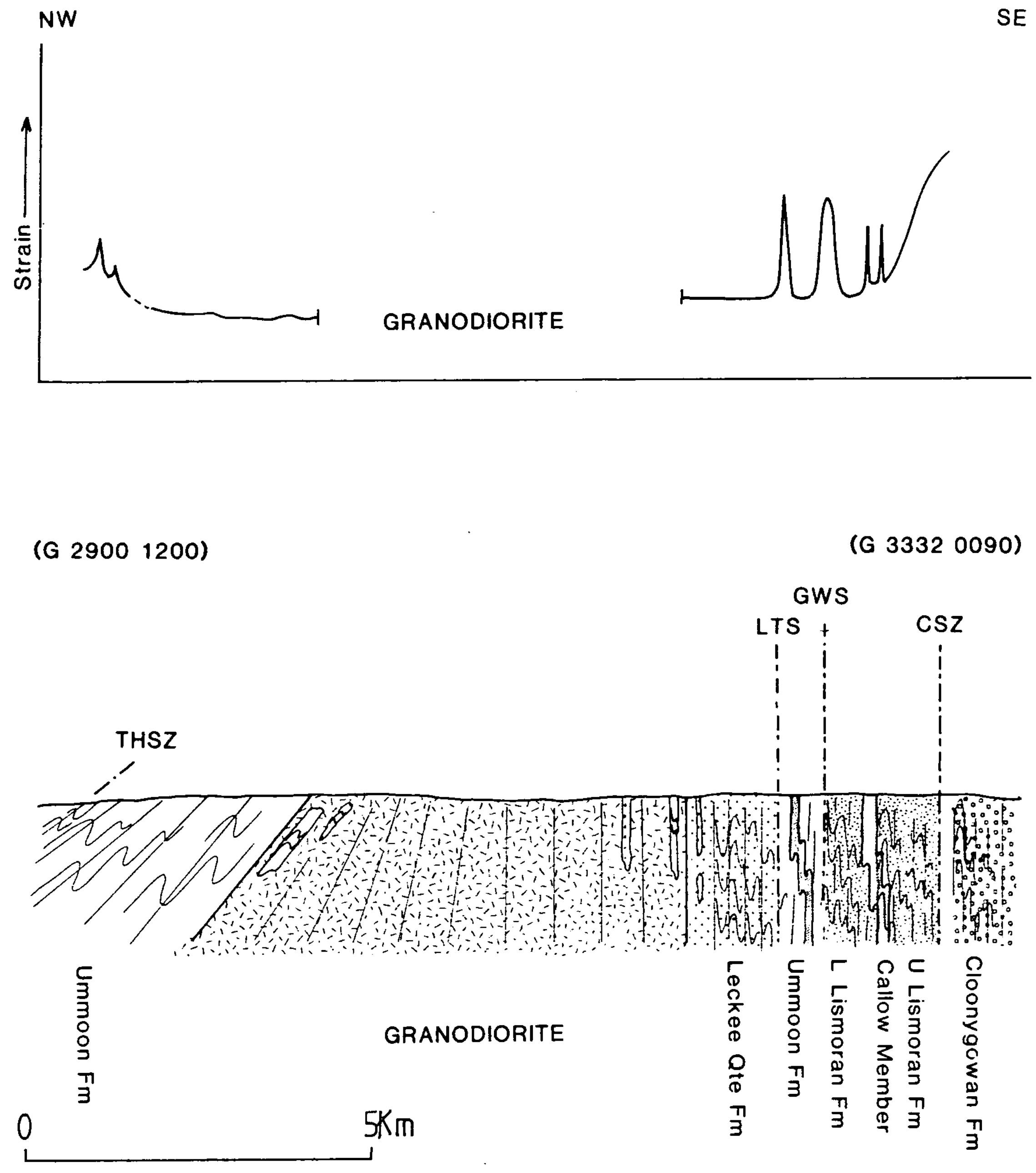
While the present study of the geometry and correlation of the deformation chronologies of both units supports the conclusion of Hutton and Dewey, (1986) that the D3 sinistral shearing affects both sets of rocks, the present work also suggests that the contact may have had an important earlier tectonic history (section 4.8). Discussion that follows in this section is limited to the features of the D3 CSZ structure. The evidence for a pre–D3 history of the tectonic contact between the Clooygowan Formation and the Ox Mountains Succession is examined in section 4.8. The detailed correlation of the deformation chronologies of both units is discussed in section 4.10.

The CSZ produced similar microstructural and geometrical effects in both the Cloonygowan Formation and the Ox Mountains Succession. In the former unit D3 strain increases towards the unexposed contact with the Ox Mountains Succession to the northwest (see figure 4.6). This is reflected by a tightening of F3 folds, decrease in the spacing of the pressure solution fabric and, in the final 10m of exposure, the development of sinistral shear bands. In thin section the S3 foliation is defined by a strong preferred alignment of flattened and ribboned detrital quartz grains set in a fine-grained muscovite/chlorite matrix. The detrital grains show a maximum x:y ratio of 10:1, with X orientated parallel to the strong gently plunging stretching lineation. This intense foliation is cut by sinistral shear bands defined by chlorite and ultra fine-grained (<0.01mm), dynamically recrystallized, quartz sub-grains (see section 4.8).

A similar increase in D3 strain is experienced by the Upper Lismoran Formation to the northwest. A gradual increase in strain to the southeast over 1000m is reflected by the tightening and eventual transposition of F3 folds, the decrease in spacing of S3 and intensification in shear band development as the contact is approached. The microstructural effects (Plate 4.17–4.18) mirror those in the adjacent Cloonygowan Formation. The intense S3 foliation is defined by a strong preferred crystallographic and dimensional orientation of quartz grains. The grains have seriate boundaries and contain subgrains which have developed in response to dynamic recrystallization. Large pressure shadows infilled with quartz, muscovite and chlorite occur around undeformed feldspar grains. Biotite is absent, indicating that the localized retrograde metamorphism associated with the CSZ occured under chlorite zone conditions. Shear bands occur on a variety of scales with a spacing that varies from <1cm to 50cm but most commonly 1-2 cm. These cut and anastamose into the pre-existing S3 foliation and consistently indicate sinistral shear. The shear bands are defined by narrow (maximum 1mm, average <0.1mm) zones of very fine grained muscovite/chlorite and quartz sub-grains < 0.01 mm. The change in metamorphic grade across the structure

FIG 4.6

HIGHLY SCHEMATIC STRAIN PROFILE ACROSS THE CENTRAL OX MOUNTAINS METASEDIMENTS



Note: no strain markers are present, estimate of strain is therefore based upon tightness of F3 folds, spacing of S3 and the development of shear bands. The effects of rheological variation due to across strike variation in lithology cannot be quantified

Plate 4.17 Thin section of the Upper Lismoran Formation on the southeast side of the Callow Shear Zone, adjacent to the Cloonygowan Formation, cut perpendicular to the shallowly plunging stretching lineation, illustrating the intense S3 fabric (G 3250 0231).

Plate 4.18 Thin section of the same specimen as above, cut parallel to the gently plunging stretching lineation, illustrating the development of sinistral shear bands that modify the S3 foliation. Sub–grains of quartz are extensively developed.

Field of view 14mm

Field of view 14mm

indicates that there is significant vertical displacement across it. This vertical displacement must have occurred some time after D3 as the rocks of the Ox Mountains Succession were still at the amphibolite facies for some time after D3 (MP3 staurolite overgrows S3). It is considered that this vertical displacement occurred along a late steeply inclined fault that is located in the unexposed ground that separates the Cloonygowan Formation and the Ox Mountains Succession at Callow (see section 4.8).

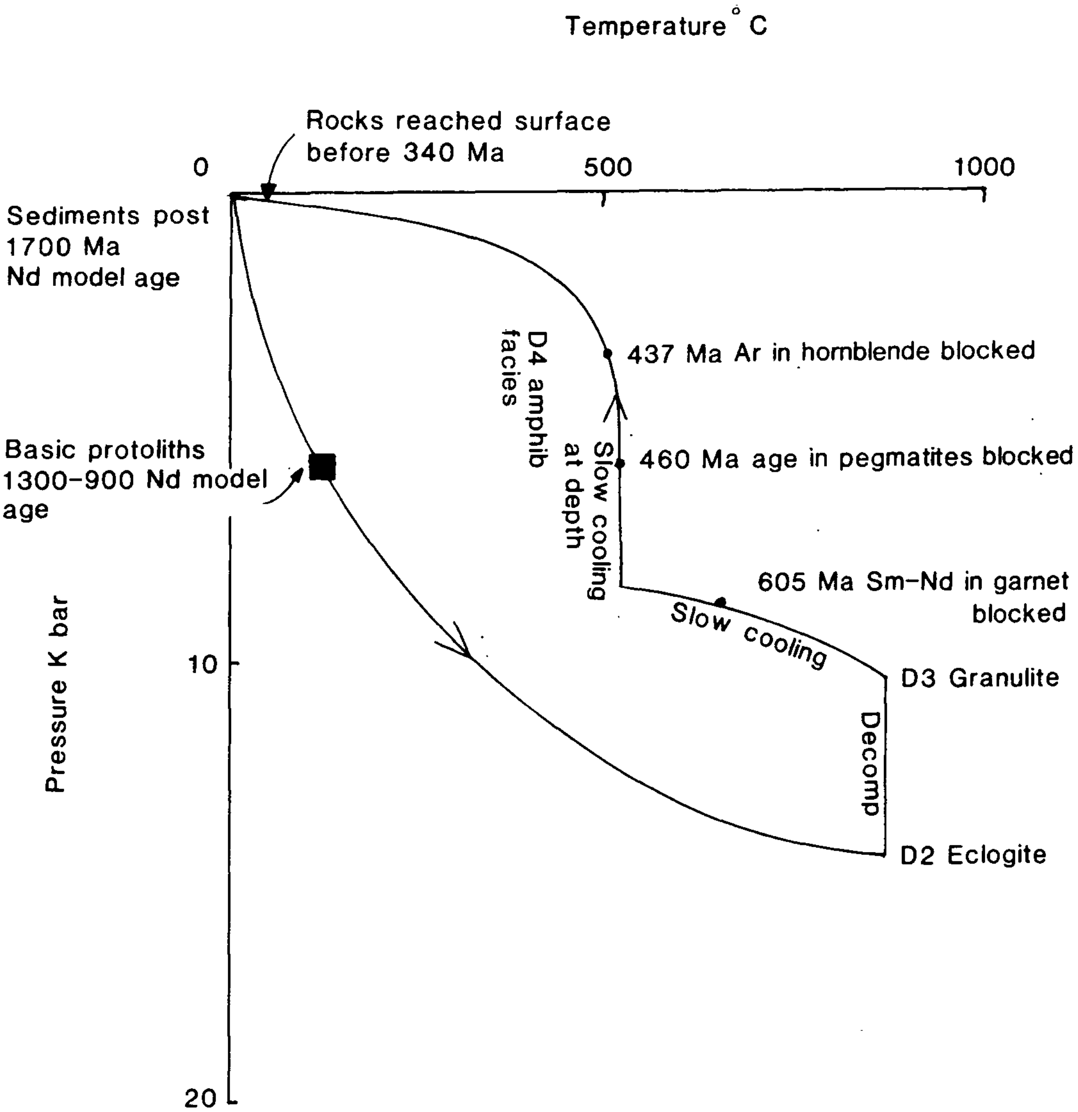
4.3.5.4 The North Ox Mountains Slide (NOXS): The nature of the contact between the granulite facies psammites of the northeast Ox Mountains and the amphibolite facies metasediments of the central Ox Mountains has been the matter of debate for some time. The contact was first studied by Lemon (1971). He regarded the structure of the rocks of the northeast Ox Mountains as Moinian, and the more variable lithologies of the central Ox Mountains as Dalradian. Lemon recorded signs of 'granulation and mylonitisation' at the boundary between the two units, which led him to suggest that the Dalradian was thrust northeastwards over the Moinian.

Phillips et al. (1975) noted that following metamorphism and deformation under granulite facies conditions the psammites of the northeast Ox Mountains were retrogressed to amphibolite facies conditions. They described an intense northeast–southwest trending flattening fabric (D2) which was associated with this retrogression and which produced a 'cataclastic' texture in some rocks. This fabric is orientated at right angles to the strike of the unretrogressed granulite facies rocks. They did not however describe the nature of the contact between the two areas of differing metamorphic history. Andrews et al. (1978) regarded the boundary between the central and northeast Ox Mountains as a high angle brittle fault, (The Ladies Brae Fault) whose displacement is essentially vertical with a small dextral component.

Yardley et al. (1979) note that the contact between the granulites of the northeast Ox Mountains and the Barrovian kyanite zone metamorphism of the central Ox Mountains is analogous to that in S. Donegal where the Dalradian is thrust over the Moinian, along the L. Derg Slide. They proposed that within the northeast Ox Mountains the 'cataclastic' deformation at amphibolite facies overprinted earlier granulite facies metamorphism. The earlier metamorphism was shown to occur at (800°C, >10kbar), significantly greater than that which affected the staurolite-kyanite schists of the central Ox Mountains (600 °C, 6-7kbar), and concluded that there was no metamorphic gradation between them. They stated that following the granulite facies metamorphism that affected the northeast Ox Mountains, these rocks were juxtaposed against the Ox Mountains Succession rocks of the southwest and central Ox Mountains which were at this stage at amphibolite facies conditions. They regarded the greenschist facies retrogression to postdate this juxtaposition.

Max et al. (1984) emphasised the difference between the basement–Dalradian contact at Lough Derg, which they regard as a D2 amphibolite facies slide and that between the northeast and central Ox Mountains Dalradian which they state is a fault. Sanders et al. (1987), examined the metamorphism of the northeast Ox Mountains in detail and noted that the high pressure granulite facies gneisses had undergone extensive Caledonian retrogression at amphibolite facies that increases in intensity towards the contact with the Dalradian of the central Ox Mountains. A similar affect was noted between the basement and the Dalradian at Lough Derg. High pressure granulite facies mineralogies are spectacularly well preserved at Slishwood, approximately 25km from the contact with the central Ox Mountains (Sanders et al., 1987). These authors also considered from probe work and textural studies of garnets in eclogites immediately north of the contact with the central Ox Mountains that temperatures here were slightly lower (700° C) than at Slishwood, (Figure 4.7). This indicates that the metamorphic grade of the northeast Ox Mountains varies from southwest to northeast (Sanders pers

FIG 4.7



PROPOSED P-T-t EVOLUTION OF THE NORTHEAST OX MOUNTAINS
(After Sanders et al. 1987)

comm, 1989). Such a lateral variation in metamorphic grade would be consistent with the northeast Ox Mountains - Lough Derg Inlier granulites representing a broad antiformal structure with the highest grades, reflecting the greatest depths, exposed towards the centre (Figure 4.8).

The geometry and kinematics of the high pressure granulite facies rocks of the northeast Ox Mountains have not been studied in detail. The data presented below is derived from the brief reconnaissance studies of the present author and the published account of Lemon (1971) and Sanders (1987). These data indicate that the contact between the northeast and central Ox Mountains is an amphibolite facies tectonic slide that produces planar high strain fabrics and retrogressive metamorphism in the psammites of the northeast Ox Mountains adjacent to the contact, and that the contact has subsequently been modified locally by high angle brittle faults. In this section the structural data obtained from the area immediately north and south of the contact between the northeast Ox Mountains and central Ox Mountains are discussed and the nature of the contact evaluated. Finally the kinematic data and a model for the structural development of this contact are introduced, although developed further in chapter 5.

The rocks of the Ox Mountains Succession are in closest proximity to those of the northeast Ox Mountains in an area known as the Ladies Brae, which is a wide northwest–southeast trending valley transecting the inlier. In most of the Ladies Brae area exposure is poor and the Ummoon Formation of the Ox Mountains Succession is separated from granulite facies psammites of the northeast Ox Mountains by a large area of blanket peat. There is no topographic feature that can be used to locate the contact, its position shown in figure 2.4 is therefore poorly constrained.

Immediately southwest of the contact with the northeast Ox Mountains the S3 foliation is shallowly inclined to <30°) to the west-southwest. Mesoscopic F3 folds, with tight-isoclinal geometries, wavelengths generally in the range 1cm–1m

Figure 4.8 Schematic development of the North Ox Mountains Slide.

A. Pre–caledonian metamorphism of the northeast Ox Mountains, under Eclogite–Granulite facies conditions, reflecting depths of approximately 35 km (Sanders, 1987).

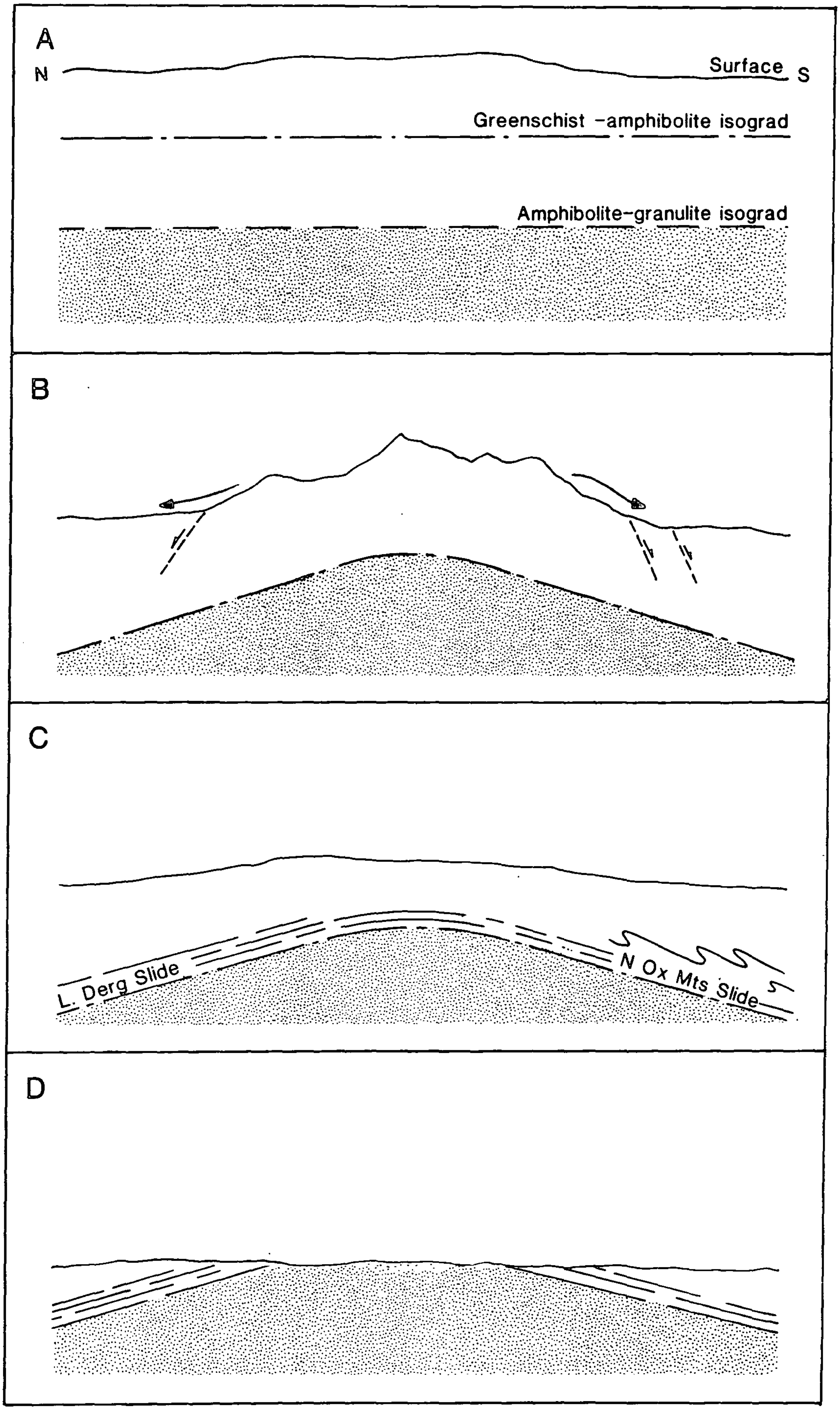
B. Erosion of uplifted high grade material previously deformed and metamorphosed during the Pre–caledonian accompanied by deposition of the Dalradian. The boundary between the Dalradian and the basement may at this stage be represented either by a faulted contact or an unconformity.

C. Caledonian metamorphism at amphibolite facies conditions. Folding in the central and southwest Ox Mountains, reactivation of the basement–cover contact. Retrogressive greenschist facies metamorphism affects the Pre–caledonian rocks, but diminishes northwards, granulite facies features are therefore preserved at Slishwood.

D. Uplift and eventual exhumation of the granulite facies rocks at Slishwood.

Note : Support for the lateral variation in grade of the basement rocks of the northeast Ox Mountains is based on the recognition (Sanders, 1987) that eclogites immediately north of the north Ox Mountains Slide contain garnets that are inclusion rich and are distinct from those that occur at Slishwood that reflect higher temperatures.

FIG 4.8


A
N
Surface
S
Greenschist -amphibolite isograd
Amphibolite-granulite isograd
B
C
L. Derg Slide
N Ox Mts Slide
D

and amplitudes between 3cm and 3m have hinges that are orientated parallel to a well developed stretching lineation with a mean orientation of 11/239 (Figure 4.5). They exhibit variable vergence.

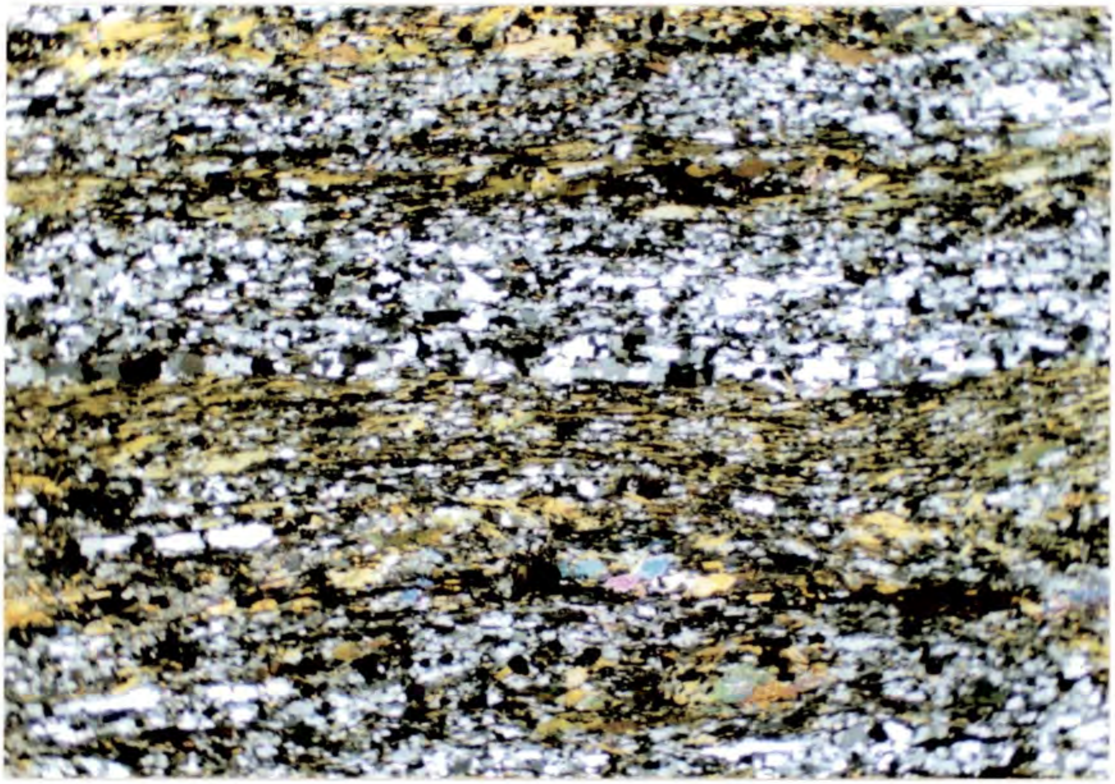
Detailed thin section examination supports field observations that D3 strain increases towards the contact with the Ox Mountains psammites (Plate 4.19). At Carrownaskeagh (G 5385 2450), 1350m from the contact the S3 fabric, which is defined by layers of muscovite and biotite (now largely retrogressed to chlorite) separated by layers of quartz and feldspar, is overgrown by MP3 staurolite porphyroblasts up to 14mm long. This indicates that metamorphic conditions during slide development were at least in the staurolite zone of the amphibolite facies (Plate 4.10). Moving north towards the contact intensification in the fabric is reflected by a decrease in the spacing of S3, and the progressive tightening of F3 folds, until they consistently display an isoclinal geometry. Post tectonic recrystallization has resulted in the re–equillibration of quartz crystals, which are now equidimensional and show no preferred crystallographic orientation. It is therefore extremely difficult to assess the intensity of D3 prior to recrystallization using textural criteria. MP3 chlorite has locally replaced MS3 biotite and garnet, indicating that retrogression followed the juxtaposition of the two units.

Limited kinematic data are preserved in the rocks immediately south of the contact. Rare occurrances of shear bands indicate overshear to the south. These data are considered to reflect the final movement on the structure, which was a minor extensional component to the south.

In the area immediately north of the contact an intense penetrative fabric is developed in the psammites of the northeast Ox. The fabric is generally either sub-horizontal or gently inclined to the southwest and has an associated strongly developed stretching lineation with a mean orientation (14/230). No folds were observed associated with this fabric. The orientation of the fabric and stretching

Plate 4.19 Thin section of the Ummoon Formation (Newantrim Member) 100 m south of the north Ox Mountains Slide, illustrating high D3 strains.

Plate 4.20 Large scale shear sense indicators developed in the granulite facies psammites of the northeast Ox Mountains, Scalpnacapple (G 6417 2772), 15km north of the north Ox Mountains Slide, which indicate north directed overshear.



Field of view 14mm

lineation is indistinguishable from that in the Dalradian of the central Ox Mountains immediately to the south, and consistently indicate that the rocks of the northeast Ox Mountains lie structurally below those of the central Ox Mountains.

Kinematic data including large scale shear sense indicators are exposed at Scalpnacapple (G 6417 2772) approximately 15km north of the contact with the amphibolite facies rocks of the Ox Mountains succession. At this locality the gently inclined mylonitic fabric, which has a northeast–southwest trending stretching lineation, is wrapped around pods of amphibolite and psammite. The internal, pre–mylonite fabric preserved in the pods is orientated at a high angle to the external fabric and consistently indicates a top to the north sense of shear (Plate 4.20). These amphibolites locally preserve clinopyroxene, which has now been largely replaced by amphibole. Sanders (1987) has obtained a step heat age of 437±2Ma. from analysis of these amphiboles which implies that the age of the deformation and metamorphism is of Caledonian age.

The granulite facies rocks of the northeast Ox Mountains and the amphibolite facies rocks of the central and southwest Ox Mountains are at their closest proximity in the stream section at Carrownamaddoo, at the northwestern end of the Ladies Brae (G 5325 2890). Garnetiferous semi–pelites and psammites of the Ummoon Formation are exposed underneath the bridge at (G 5235 2850). These exposures structurally overlie pale weathering granulite facies feldspathic pasmmites of the northeast Ox Mountains granulites that outcrop immediately downstream.

The fabric that is developed within these psammites is indistinguishable in orientation from that of the amphibolite facies metasediments of the Ox Mountains Succession immediately to the south. The intensification of the S3 fabric in the Ummoon formation northwards towards the contact suggests that the contact between the two units is a tectonic slide of D3 age. Metamorphic data (Plate

4.10) suggest that movement occurred under amphibolite facies conditions prior to the growth of MP3 staurolite porphyroblasts.

Further downstream it becomes difficult to distinguish between lithologies of the Ox Mountains succession and those of the northeast Ox Mountains granulites. Rocks consisting of a high proportion of semi-pelite, that resemble those of the Ummoon Formation are interbanded on a scale of 1–5m with homogeneous feldspathic psammites that are unequivocal northeast Ox Mountains granulites. The boundaries between these layers lie parallel to the strong penetrative S3 foliation in both sets of rocks. The present author considers that in the immediate vicinity of the contact, the rocks of the northeast Ox Mountains and the Ox Mountains succession are tectonically interleaved or imbricated. This is a common feature of slides, (Hutton, 1979b). Such interbanding of rock units results in extreme difficulty in locating the precise contact between the two units as it is not simply one contact, but in fact a number of parallel tectonic contacts.

Moving further downstream from this interbanded zone, gently dipping homogeneous psammites of the northeast Ox Mountains are continuously exposed between (G 5315 2954) and the margin of the inlier at (G 5315 2970).

The contact between the granulites of the northeast Ox Mountains and the amphibolite facies metasediments of the Ox Mountains succession is therefore considered to be a D3 tectonic slide, referred to here as the North Ox Mountains Slide. This contact is affected by late stage brittle faulting accompanied by localized brecciation and cataclasis. The absence of a major topographic feature and common orientation of fabrics in both units that increase in intensity towards the contact between them, suggests that late faulting has a relatively minor effect upon the structure (cf. Max et al., 1984).

It is unclear whether the slide was originally an unconformity or a tectonic (faulted) contact (Figure 4.8), that has subsequently been reactivated. The present situation, where the Ummoon Formation is juxtaposed against the Ox Mountains granulites, indicates that in the absence of extreme facies variation, a significant part of the Ummoon Formation and all of the Leckee Quartzitic Formation have been cut out. Given that the angular discordance between the lithological banding in the Ox Mountains Succession and the foliation in the slide zone is regionally small, this suggests that the movement on the slide was large.

4.3.6 Contact with the Ox Mountains Granodiorite: evidence for the timing of emplacement

The contact between the Ox Mountains Granodiorite (section 2.5.1) and the metasedimentary envelope is exposed along the southeast side of the inlier between Sallagher (M 1670 9861) and Attimachugh (G 3240 0620). It is a steeply inclined and intensely deformed sheeted contact zone in which sheets of granodiorite increase in number and thickness towards the continuous body of granodiorite. Critical evidence for the timing of emplacement of the pluton is widespread but is best exposed at two main localities. Firstly in the Ummoon Formation at Ummoon (G 2710 0012), sheets of granodiorite cut across tight-isoclinal F3 folds, indicating that intrusion postdated the development of F3 folds within the inlier (Plate 4.21). The steeply inclined SC fabric developed within the sheets has an orientation that is indistinguishable from that in the metasediments, where it is axial planar to F3 folds (Plate 4.22). This relationship indicates that the emplacement occurred synchronously with the fabric development in the country rocks following initial F3 fold development. Secondly at (G 0427 3057) on the Foxford-Swinford road, large sillimanite porphyroblasts up to 5cm in length are developed in sheets of semi-pelite and pelite within the Leckee Quartzitic Formation only in the immediate vicinity of the granodiorite. These porphyroblasts are assumed therefore to be due to the thermal metamorphism of the pluton. The porphyroblasts both overgrow and are locally deformed by the SC fabric in the country rocks. This suggests that intrusion was syn-kinematic with respect to the

Plate 4.21 Tight to isoclinal, northwest verging F3 folds developed in the Leckee Quartzitic formation, cut by sheets of granodiorite, Ummoon (G 2710 0012).

Plate 4.22 Granite vein intruding psammite of the Leckee Quartzitic Formation, deformed by the subvertical S3 foliation that is co–planar to the S3 foliation that is developed in the metasediments, Ummoon (G 2710 0014).

deformation of the country rocks and that deformation overlapped cooling of the pluton and development of the thermal aureole. This conclusion is supported by evidence from the rest of the pluton and is consistent with the diagnostic criteria for syn-kinematic plutons proposed by Hutton (1988).

4.3.7 D4

D4 produced a set of angular, open to close folds that frequently show a conjugate geometry. (Plate 4.23) These structures can be identified throughout the inlier but find their greatest development in area 1c where the amplitude of F4 folds reaches a maximum of 100m, and commonly exceeds 10m, which is an order of magnitude greater than observed in the remainder of the inlier. The axial planes of these structures are always inclined at a high angle to S3. Their orientation falls into two groups which reflects the commonly conjugate nature of F4 folds. F4 fold axes consistently lie parallel with or at a small angle to the well developed X3 lineation which represents a strong anisotropy. The orientation of F4 fold axes therefore shows a similar distribution to F3 folds. Quartz crystals show undulose extinction and an occasional weak preferred dimensional orientation parallelto the axial planes of the folds. MS3 muscovite and chlorite are bent by D4, but show no recrystallization, which indicates that they developed at very low metamorphic grade. The geometry of D4 structures suggests that they represent vertical shortening or strike-perpendicular extension. The probable kinematic significance of these structures is discussed in section 5.4.

4.3.8 Contact relationships with the Lough Easky adamellite : evidence for the timing of emplacement.

The contact between the Lough Easky Adamellite and the metasediments of the Ummoon Formation is very poorly exposed. Exposure of Adamellite veins at Lough Callow (G 4481 2025) that cut across F4 folds (Plate 4.24) indicates

Plate 4.23 F4 folds, with gently inclined axial planes, Ummoon Formation, Castlerock (G 4407 1510).

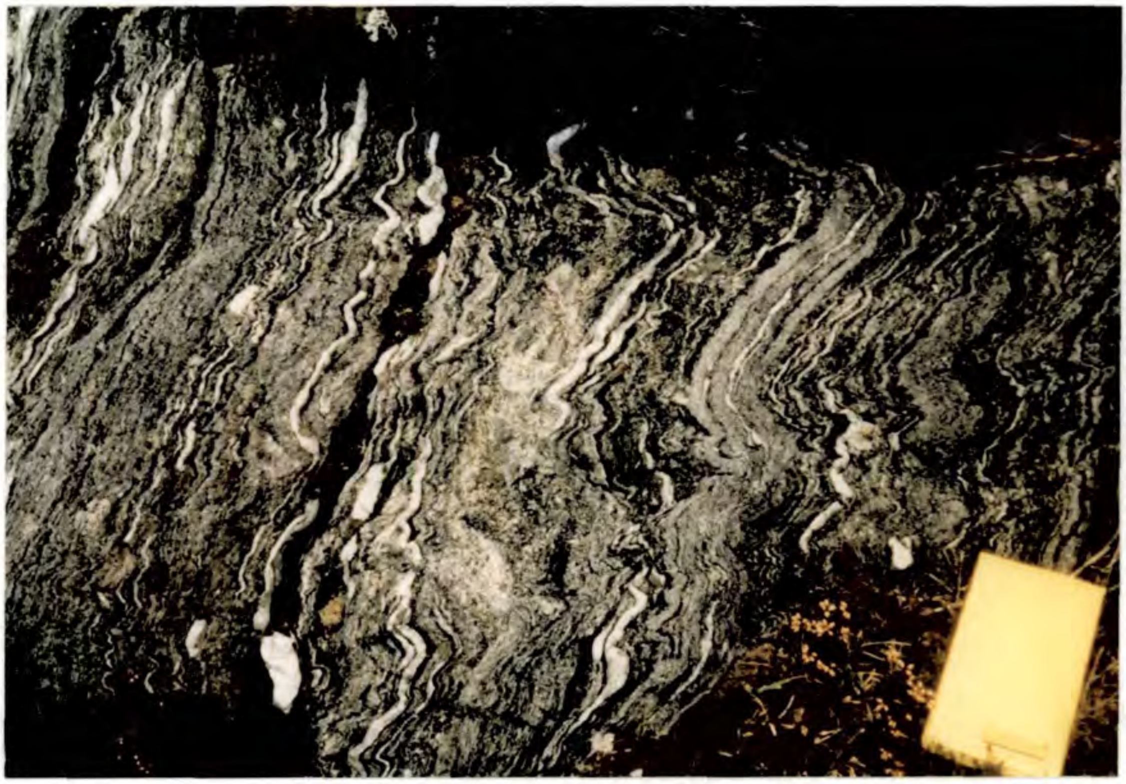


Plate 4.24 F4 folds cut by a steeply inclined vein of adamellite, Lough Callow, near Lough Easky (G 4487 2025).

that emplacement occurred post-D4. This vein relationship is confirmed by thin section examination which reveals that large andalusite porphyroblasts associated with the aureole of this pluton randomly overgrow F4 folds.

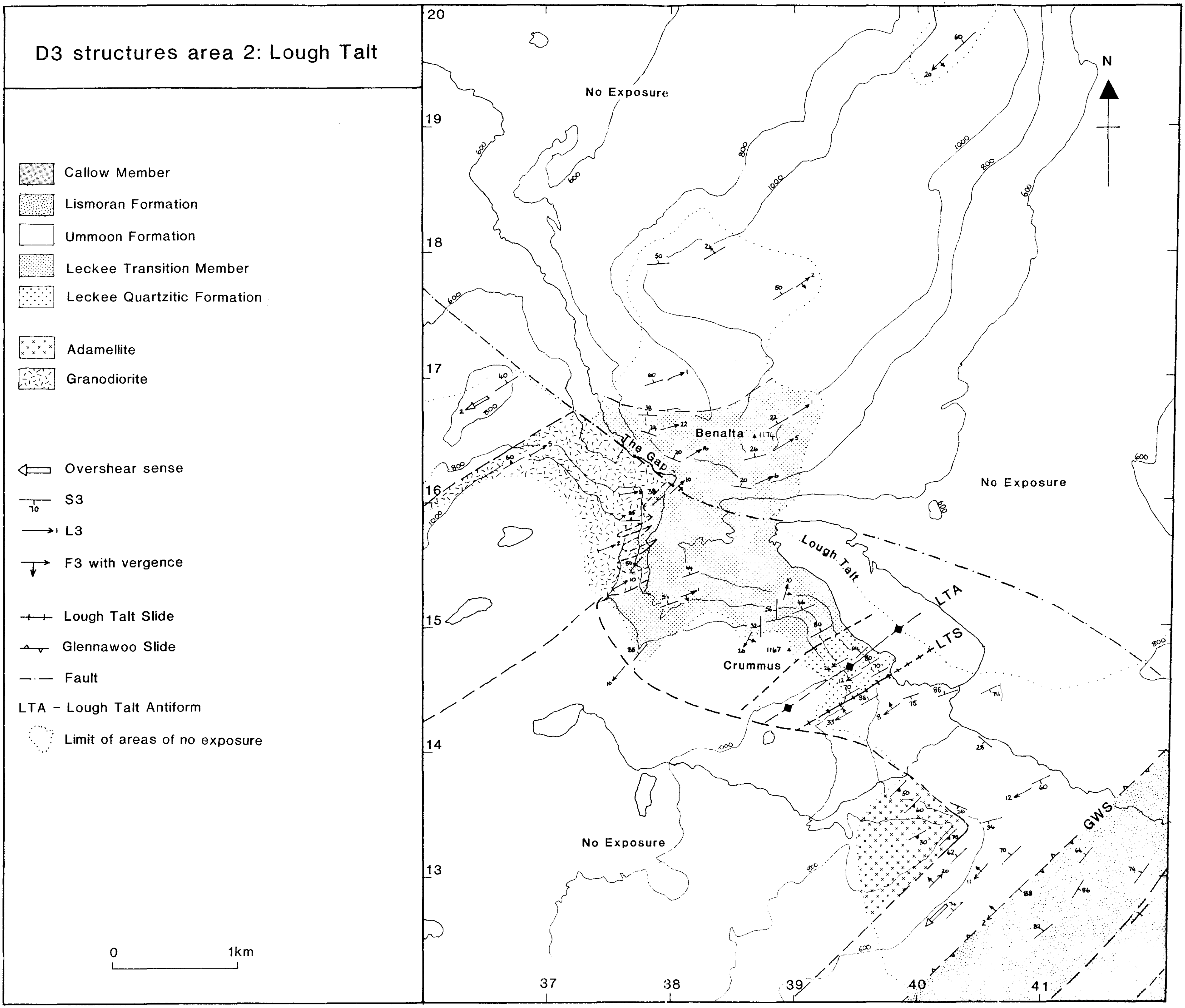
4.4 AREA 2 : LOUGH TALT

This area, which is located immediately north of the Ox Mountains Granodiorite, is of critical importance as it provides the only link between the structure of the area northwest of the Ox Mountains Granodiorite and the area to the southeast, and also exposes the contact between the Lough Talt Adamellite and the metasediments. Structurally the area is complex. It includes the core of a major F3 antiform that controls the outcrop pattern in the metasedimentary envelope to the granodiorite (Figure 4.9). In addition the Lough Talt Slide is present in the southeast of the area and F2 folds also occur here.

The importance of the area was recognised by previous researchers in the central Ox Mountains. Both Taylor (1968) and Andrews et al. (1978) attempted to draw cross-sections through the inlier here, but experienced great difficulty in relating the structural geometry that they observed to their interpretation of the stratigraphy. In this section the problem of the stratigraphic identitiy of the metasediments is reviewed (see section 2.2 for details) before the structural geometry of the area is described in detail. A model is then proposed to account for the relationship between the structural and stratigraphic relationships observed. Finally the structural models of the previous workers are reviewed critically and conclusions drawn (Figure 4.10).

In common with the remainder of the inlier the boundaries between stratigraphic units are orientated parallel to the S3 fabric, which in the case of the Lough Talt area fans from an orientation of 226/38 at (G 3790 1600) to 058/90 at (G 3976 1460) (Figure 4.10). The detailed stratigraphic correlation with

FIG 4.9



D3 structures area 2: Lough Talt

other areas in the Ox Mountains discussed in section 2.2 has resulted in the re-interpretation of the identity of some of the stratigraphic units. Briefly, while there is agreement with previous workers that the pelite, semi-pelite and psammite south of Lough Talt and north of the contact with the Lough Talt Adamellite represents the Ummoon Formation, and that the pure, white weathering homogeneous orthoquartzite exposed immediately northwest of this represents the Leckee Quartzitic Formation, there is disagreement about the identity of all of the stratigraphy exposed to the northwest of this.

These metasediments have been identified as Attymass Group, Ummoon Formation and Leckee Quartzitic Formation by Taylor (1968), and as Attymass Group and Leckee Quartzitic Formation by Andrews et al. (1978). The interpretation of the present author of these metasediments as a transition member or lateral facies variation within the Leckee Quartzitic Formation greatly simplifies the stratigraphy and is discussed in section 2.2.1.

Correct identification of this material is critical to the construction of a stratigraphic section through the area. The differing stratigraphic interpretations of the area led previous workers to propose a number of complex structural models for the area which are not supported by field structural evidence. The cross-section presented by Taylor is inconsistent with his map (his Fig 7 pp 577, present figure 4.10a). It appears however that he regards the LTS as folded around the major antiformal axis that is located on Crummus (G 3895 1480). No evidence has been observed to support this, and the present author agrees with Andrews et al. that there is no observable incease in strain northeast of Crummus at the slide location described by Taylor.

Andrews et al. (1978), noted that the vergence of F3 folds changed close to the summit of Crummus. However F3 folds alone were unable to account for the stratigraphic configuration that they recorded. They therefore 'postulated' the

Comparison of cross sections through Lough Talt

⁙ Cloonygowan Fm.
▒ Lismoran Fm.
☐ Ummoon Fm.
░ Leckee Quartzitic Fm.
▧ Attymass Gp.

Taylor (1969)

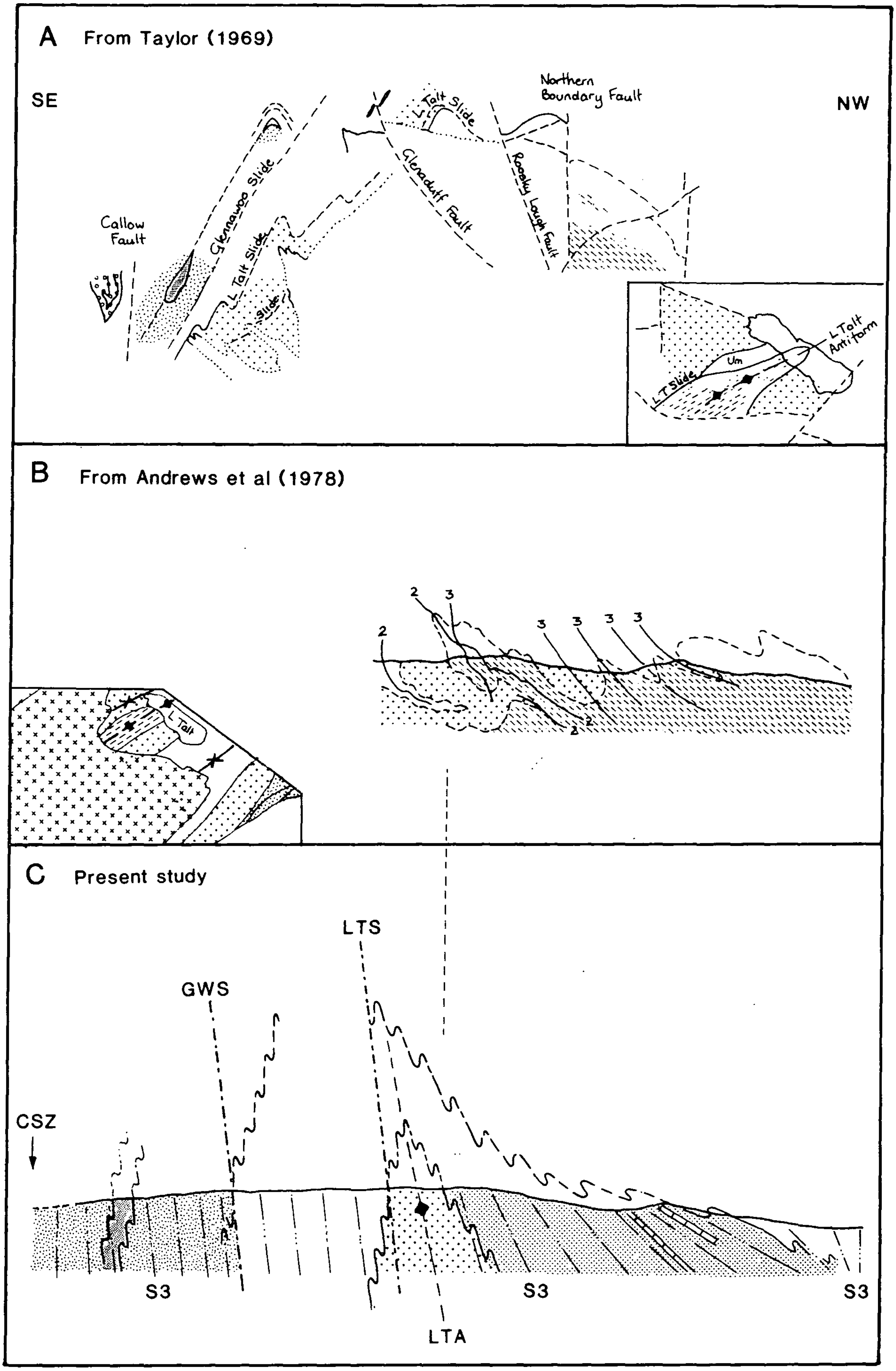
▒ Lismoran Fm.
☐ Ummoon Fm.
░ Leckee Quartzitic Fm.
▧ Attymass Gp.

Andrews et al.(1978)

▒ Lismoran Fm.
☐ Ummoon Fm.
▓ Leckee Transition Member
░ Leckee Quartzic Fm.

Present study

FIG 4.10



A From Taylor (1969)

B From Andrews et al (1978)

C Present study

existence of a number of F2 isoclinal folds to account for the observed stratigraphy, but did not cite any structural evidence in support of this.

The present interpretation, based on 1:10.560 mapping of the area (Figure 4.10c) is that the variable lithology exposed to the west of the unequivocal Leckee Quartzitic Formation represents the Leckee Tansition Member, a transitional unit between the quartzite and the Ummoon Formation. The unit does not appear on the southeast side of the Leckee Quartzitic Formation as it is cut out by the Lough Talt Slide. This model satisfactorily accounts for the difference in lithostratigraphy on either side of the Leckee Quartzitic Formation which itself lies in the core of a major F3 antiform, and therefore structurally underlies the exposures on the northwest and southeast sides of it.

4.4.1 D1–D2

D2 is expressed as a composite S1–S2 fabric that is orientated parallel to lithological banding. S2 is transposed by S3 in most of the area but can be positively identified in the hinges of F3 folds, where it is a penetrative fabric consisting of quartz, muscovite and biotite overgrown by MP2 garnet and feldspar. In common with other areas in the inlier F2 folds are extremely rare and isolated occurrances are only observed in two localities, both within the Leckee Quartzitic Formation where it has resisted transposition by D3 (Plates 4.25–4.26). At (G 3967 1465) on the north facing slopes of Crummus a series of F2 folds can be observed refolded by later F3 structures. The F2 axes plunge subvertically, have axial planes that are orientated northeast-southwest and consistently verge to the northwest. This orientation is totally inconsistent with the regional geometry of S2 which, if the effects of F3 folds are removed, is shallowly inclined elsewhere in the inlier. In addition the relationships observed at Lough Talt, where F2 and F3 folds axes make an angle of 90° to each other are inconsistent with data from the rest of the Inlier where F2 and F3 folds are co-axial and interfere giving type

Plate 4.25 West–northwest verging F2 fold, horizontal surface, viewed looking north–northeast, Leckee Quartzitic Formation, Lough Talt (G 3967 1465).

Plate 4.26 North–northwest verging F3 folds (same outcrop as 4.25), vertical surface, viewed looking south–southwest.

three interference patterns. If the effects of F3 folding are removed at Crummus and the F2 folds are simply unfolded about F3, so that S2 has a horizontal attitude, the resultant D2 geometry is that of a series of recumbent folds with northwest–southeast trending axes that close towards the northeast. These data are insufficient to demonstrate a major F2 fold closure in this area. In any case stratigraphic correlations do not require it.

4.4.2 D3

The third deformation is responsible for the main fabric seen at Lough Talt, which consists of a preferred dimensional orientation of quartz, muscovite and biotite. S3 augens MP2 garnet porphyroblasts that contain MS2 inclusions of muscovite and quartz, and is itself overgrown by MP3 staurolite porphyroblasts at (G 3783 1491). This indicates that at the time of the formation of this fabric, metamorphic conditions lay within the amphibolite facies.

The fabric has a mean orientation of 243/73, but forms a cleavage fan with shallow to moderate northwest dips at Tawnyany in the northwest (G 3830 1570), gradually steepening to the vertical east of Crummus (G 3930 1494) and changing to steep southeast dips at Larbaun (G 3980 1430) (Figure 4.11). Although this fabric is intense, and associated F3 folds are tight, the deformation is less well developed than on the southeast side of the inlier, and shear bands are rarely developed. F3 folds which are developed throughout the area have an upright tight-isoclinal geometry and a mean orientation of 09/246, parallel to the well developed L3 stretching lineation. Vergence of F3 structures indicates the presence of a large, kilometric scale upright, tight antiform whose axis is located within the Leckee Quartzitic Formation on the northeast facing slopes of Crummus (G 3955 1475). This is termed the Lough Talt Antiform.

Figure 4.11 Stereographic projection of structural data : Area 2 (Lough Talt)

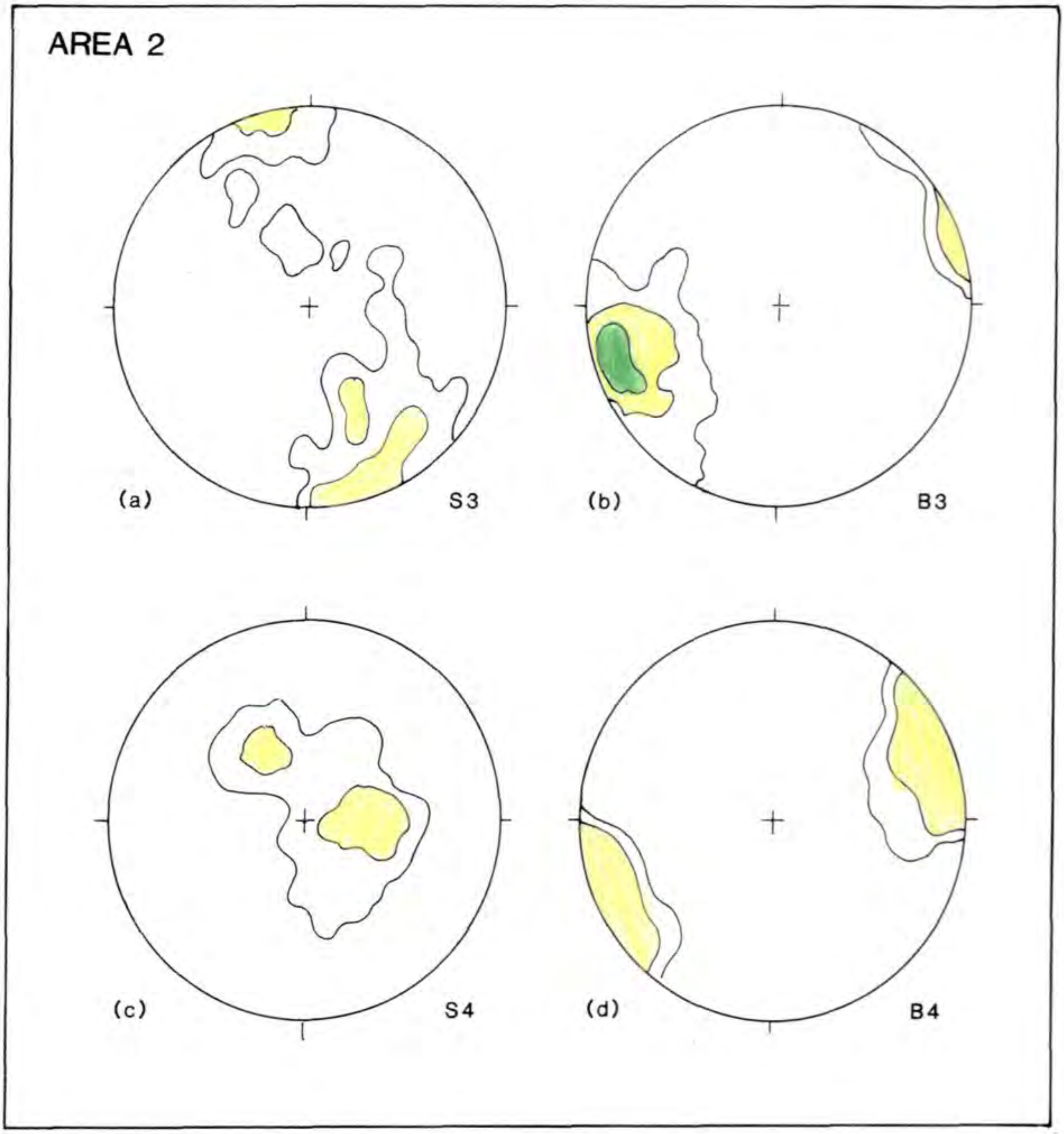
(a) S3 : mean orientation of foliation plane 243/73, (pole, 17/153), n=88

(b) $\beta3$: mean orientation of lineation 09/246, n=64

(c) S4 : mean orientation of axial plane 150/11 (pole, 79/060), n=27

(d) $\beta4$: mean orientation of lineation 13/246, n=19

Note–See Figure 4.5 for key to contour intervals.


AREA 2
(a)
S3
(b)
B3
(c)
S4
(d)
B4

4.4.3 D4

Structures related to D4 are common in the Lough Talt area and are best observed in the pelitic / semi-pelitic lithologies. As elsewhere in the inlier F4 fold axes lie parallel to the L3 stretching lineation, have a mean orientation of 13/246, and are open structures with an angular and commonly conjugate geometry. The wavelength of these structures in this area does not exceed 1m, with a corresponding amplitude of 1m. S4 axial planes are inclined at a high angle to S3 and fall into northwest and southeast dipping groups. As with other areas in the inlier the metamorphic grade at the time of the formation of D4 structures was low, MS3 micas are bent by F4 folds and quartz shows undulose extinction and a weak preferred dimensional orientation parallel to S4 axial planes.

4.4.4 The metasediment contact with the Lough Talt adamellite: evidence for the timing of emplacement

The Lough Talt Adamellite is intruded into metasediments of the Ummoon Formation in the southeast of the area. The emplacement mechanism of the pluton is examined in chapter 5. Evidence for the timing of emplacement of the pluton is widespread, the adamellite cuts across F4 folds in a number of localites on the northeast facing craggy slopes south of Lough Talt (G 4050 1305). A 100m wide contact aureole is developed in the pelitic and semi-pelitic lithologies of the Ummoon Formation, with large (maximum 3cm in length) sillimanite and biotite overgrowing a regional metamorphic assemblage of quartz, feldspar, biotite, muscovite, chlorite and garnet.

4.5 AREA 3, NORTHWEST SIDE OF THE INLIER: ATTYMASS–TAWNANEILLEEN

For the purposes of geometrical analysis of D3 structures area 3 has been divided into northeastern and southwestern sub-areas (areas 3a and 3b respectively, Figure 4.12).

Metasediments of the Ox Mountains Succession (largely the Ummoon Formation) are exposed on the northwest side of the inlier between (G 2715 0990) and (G 4450 2500) 1km north of Lough Easky. Exposure of the area is variable; it is locally good at Attymass and at Benalta (G 3866 1652) in the northern part of the area, but deteriorates towards the southwest in the low ground that is dominated by Carboniferous rocks and blanket peat. For 96 % of the strike length Carboniferous rocks form the northwest limit of exposure, however at Lough Brohly and Ballycong, rocks that probably represent part of the Raheen Barr Succession of the North Mayo Inlier Dalradian (Long and Max, 1977) separate the Ox Mountains Succession metasediments from the Carboniferous further to the northwest. South of the Gap (G 3587 1625) the Ox Mountains Granodiorite (whose contact relationships were discussed in detail in chapter 2) forms the southeast limit of exposure. At Lough Talt, north of the Gap, a large area of blanket peat cover separates the area from the rocks of area 4. The northeast and southwest limits of exposure are delimited by an area of blanket peat north of Lough Easky and Lough Conn respectively.

4.5.1 Ss

Lithological banding has been distinguished throughout the area. It is expressed as a composite Ss/S1/S2 fabric. Lithological banding has been substantially reorientated by later structures over much of the area but a shallowly dipping limestone band 10m thick, can be traced for 1.5km on the south facing slopes of Benalta (G 3866 1652). Throughout the area lithological banding dips

Figure 4.12 Stereographic projection of structural data : Area 3 (N.W. Side of the inlier)

Area 3a : Attymass

(a) S3 : mean orientation of foliation plane 232/35, (pole, 55/142), n=283

(b) $\beta 3$: mean orientation of lineation 29/283, n=124

(c) S4 : mean orientation of axial plane 352/16 (pole, 74/262), n=112

2d) $\beta 4$: mean orientation of lineation 18/260, n=81

Area 3b : Tawnaneilleen

(a) S3 : mean orientation of foliation plane 231/48 (pole, 42/141), n=119

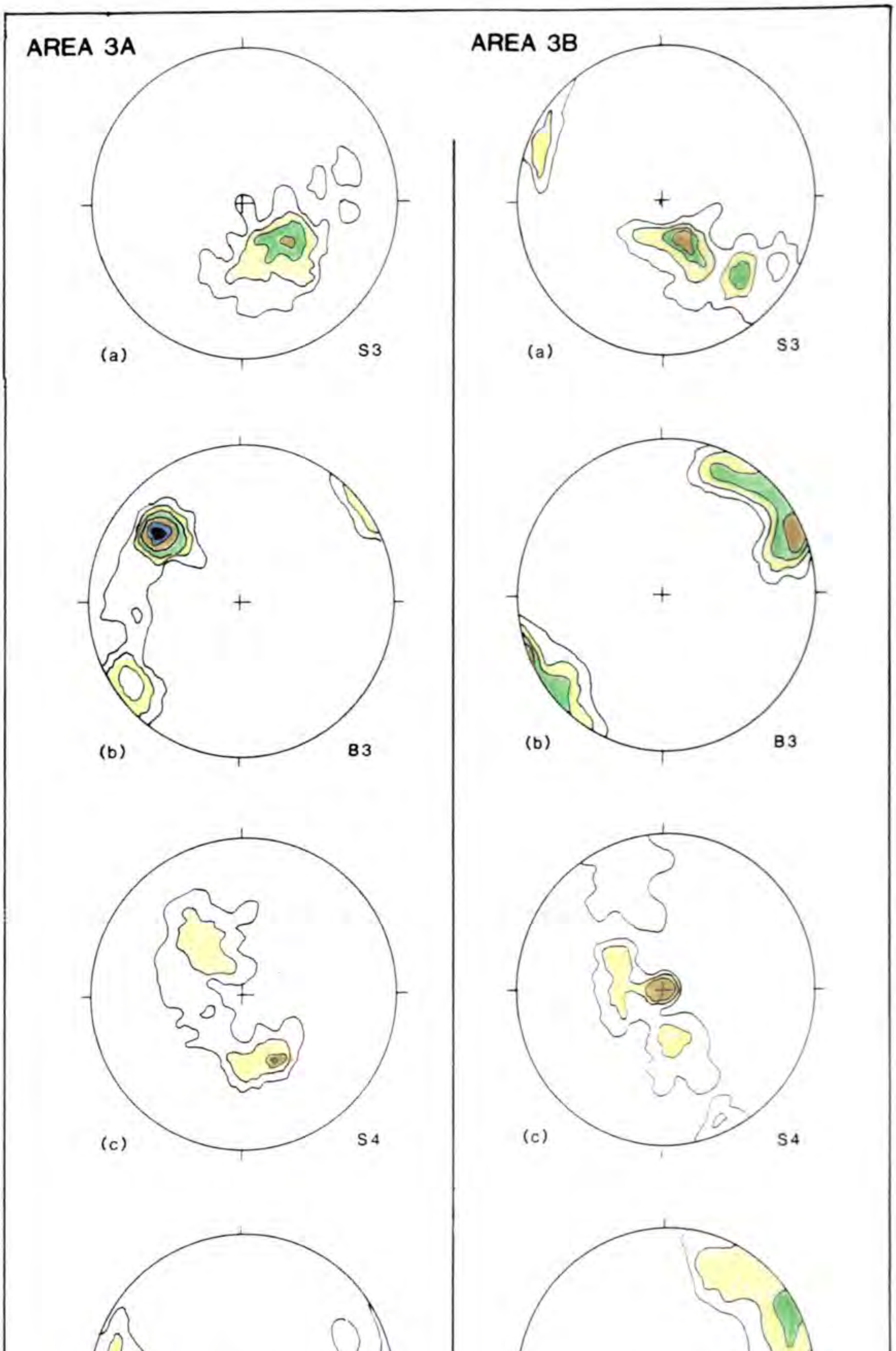
(b) $\beta 3$: mean orientation of lineation 04/052, n=99

(c) S4 : mean orientation of axial plane 041/22 (pole, 68/311), n=63.

(d) $\beta 4$: mean orientation of lineation 08/050, n=21

See Figure 4.5 for key to contour intervals.

FIG 4.12



gently (<45°) to the northwest (estimate based on the sheet dip of D3 structures). Way up criteria have not been observed in the area, correct way up is however inferred from the stratigraphic study of chapter 2.

4.5.2 D1

S1 has only been identified positively at Tawnaneilleen (G 4240 2126), west of Lough Easky where a bedding parallel fabric is folded around the hinge of an F2 fold that has subsequently been refolded by F3. At this locality S1 is expressed by a bedding-parallel penetrative alignment of muscovite and quartz. No folds of F1 age have been identified in the area.

4.5.3 D2

The penetrative S2 foliation can only be distinguished from S3 in the hinges of F3 folds, where it is represented by a biotite, muscovite, quartz (MS2) fabric, that lies parallel to lithological banding. Elsewhere S2 has been totally transposed into, and is therefore indistinguishable from, the intense S3 fabric. Neither the vergence of S2 on bedding, nor F2 fold vergence has been observed; the pre-D3 geometry can not therefore be determined. The rarity of F2 folds, and absence of stratigraphic repetitions attributable to this phase, suggests that large scale D2 structures are either absent from the area or on too large a scale to be distinguished within the outcrop area. The fact that the band of limestone exposed on Benalta (G 3866 1652) can be traced for 1.5km suggests that in this area at least, the stratigraphy has not been significantly disrupted by D2 structures. Feldspar porphyroblasts are widely developed throughout the area and are pre-D3. It is however not clear whether they represent MS2 or MP2 crystallization.

4.5.4 D3

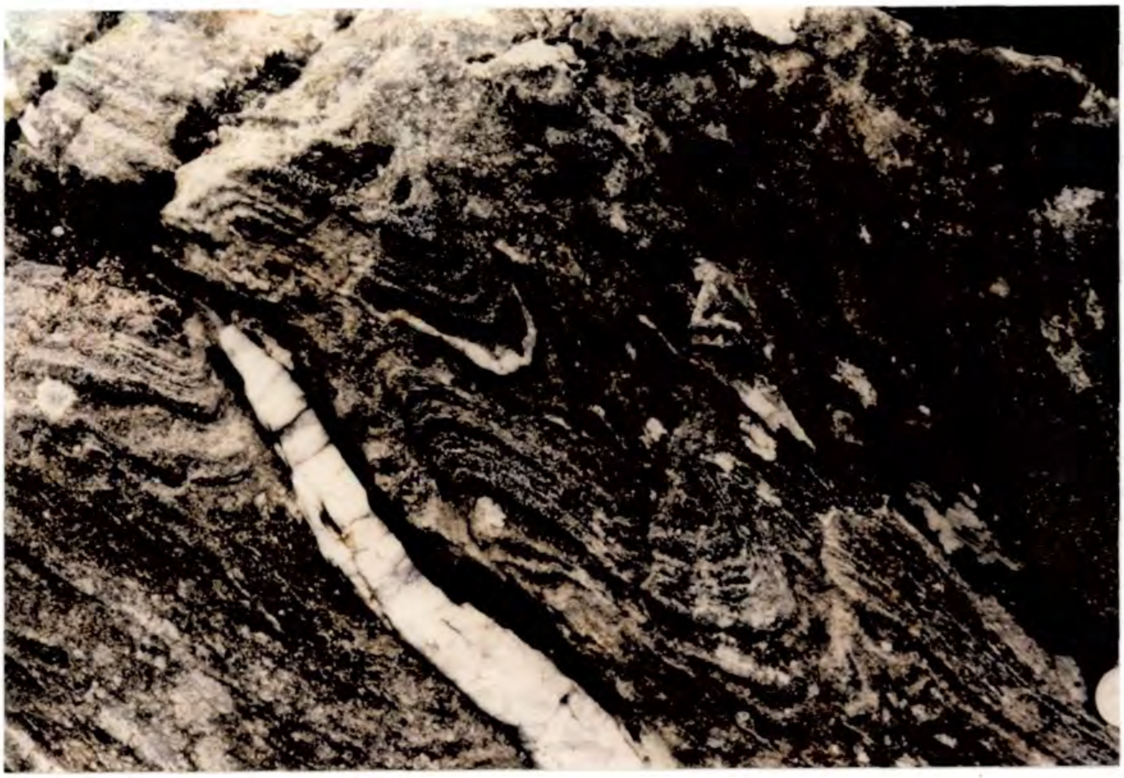
The entire area has experienced intense D3 deformation which is represented by tight-isoclinal F3 folding and development of an S3 axial planar fabric, that has almost completely transposed all earlier fabrics and is itself locally modified by SC3 shear bands. Primary evidence for the age of this fabric is rare (see section 4.6.2); therefore other criteria such as the geometry of structures, relationship to the regional metamorphism and cross-cutting relationship with the granodiorite have all been used to determine that the fabric of this area is of regional D3 age. F3 folds have a lower amplitude : wavelength ratio than those exposed in area 1. They typically have wavelengths in the range 0.1–1m and amplitudes of between 0.2 and 2m.

Although locally reorientated by later deformation at Attymass (G 3000 1200) and Benalta (G 3866 1662), S3 has a uniform strike for most of the exposure length on the NW side of the inlier. The dip of the fabric is more variable, and the mean orientation therefore changes from moderate dips at Attymass and Lough Talt (28–35°) (areas 3a, Figure 4.12) before steepening to an average dip of 68° at Tawnaeilleen (area 3b, Figure 4.12). The latter area is further to the northwest and therefore structurally higher in the sequence.

F3 folds have a tight-isoclinal geometry, and consistently verge to the southeast (towards the contact with the granodiorite,) (Plate 4.27). F3 fold axes lie parallel to the X3 stretching lineations and are plotted as β lineations (Figure 4.12). F3 folds within the area are of constant scale, and have a mean amplitude of 20cm and mean wavelength of 10cm. There is no evidence of local changes in the vergence of F3 folds that would indicate the presence of mesoscopic (> outcrop scale) folds of this age. The consistent southeast vergence of S3 with respect to SS and S2 indicates that the area lies on the northwest limb of a major F3 antiformal structure.

X3 stretching lineations plunge gently to the northwest varying from a mean orientation of 29/283 in area 3a where the effects of later deformation is most

Plate 4.27 Southeast verging, close F3 folds, developed in the Ummoon Formation, Attymass (G 3100 1182).



prominent, to 04/052 in area 3b. Their gentle plunge reflects the transcurrent nature of the deformation.

4.5.5 The Tawnwneilleen High Strain Zone

In general terms, the intensity of D3 deformation along the northwest side of the inlier is less than that which affects the southeast side of the inlier. While no tectonic slides have been identified, at Tawnaneilleen (G 4040 2100), the dip of the S3 foliation, which is normally inclined gently to the northwest on the northwest side of the inlier, locally steepens into a number of zones of increased D3 strain which can be traced for of 4km along strike before exposure is lost (Figure 2.4). As these areas of steep S3 foliation occur entirely within the Ummoon Formation it is not possible to determine the displacement across the zones. Nevertheless it is apparent that they represent a significant increase in D3 strain over the background level. This area of increased D3 strain is therefore referred to as the Tawnaneilleen High Strain Zone (THSZ).

4.5.6 Contact with the Ox Mountains Granodiorite

The sheeted contact between the metasediments of area 3 and the Ox Mountains Granodiorite is exposed sporadically between the Gap (G 3787 1625) and Lough Conn (G 2075 0608). Thin sheets of granodiorite orientated parallel to S3 intrude the metasediments and cut across F3 folds. These sheets increase in thickness and number towards the pluton the margin of which is arbitrarily defined as where the proportion of granodiorite exceeds 50%. A strong SC fabric is developed in the granodiorite, which is indistinguishable in geometry and orientation from that developed in the metasediments. Nowhere has the granodiorite contact been observed to be discordant with this fabric. Critical evidence for the timing of the emplacement of the Ox Mountains Granodiorite is observed east of Attymass at (G 3100 1133) where a gently plunging F4 fold, with a shallowly inclined axial plane deforms the contact between a thin (30cm) sheet of granodiorite and pelitic metasediments of the Ummoon Formation. This confirms that intrusion pre-dated D4.

4.5.7 D4

As in other areas F4 fold axes lie parallel with or at a small angle to the strongly developed X3 stretching lineation. The orientation of F4 therefore varies systematically with the orientation of X3. In area 3 F4 folds which have an angular geometry, are mesoscopic in scale with a wavelength of less than 1m and a corresponding amplitude of less than 1m. Their axial planes fall into two groups both of which intersect the composite SS/S1/S2/S3 fabric at a high angle, and therefore in this area dip to the northwest and southeast (Figure 4.12) and commonly interfere, forming conjugate sets. Phyllosilicate minerals are bent around the hinge of the F4 folds but are not recrystallised. Quartz shows evidence of high strain, with the development of undulose extinction and occasionally a weak preferred dimensional orientation, indicating that in common with the rest of the inlier, D4 structures formed at very low grade.

4.5.8 Contact with the metasediments of the Raheen Barr Sucession

The contact between the poorly exposed marble, quartzite, psammite and semi-pelite of the Raheen Barr Succession (Long 1977) and the Ox Mountains is not exposed. No discernable increase in strain has been observed in the Ox Mountains Succession rocks closest to the contact, and in addition, the tectonic fabrics within the Raheen Barr Succession are highly discordant with those in the Ox Mountains Succession. The most southeasterly exposure of the Raheen Barr Succession at (G 3134 1438) is extensively brecciated indicating northeast-southwest trending faulting. It is likely therefore that the contact between the two units is presently expressed as a brittle fault.

An isolated exposure of quartzite at (G 3135 1447) exhibits mylonitic textures. Unfortunately this isolated outcrop has been brecciated and reorientated by the later faulting; it is therefore not possible to determine the original orientation of the mylonitic fabric. It is not clear if this fabric results from deformation that brought the Raheen Barr Succession into tectonic contact with the Ox Mountains Succession or was developed earlier. The trend of the contact (inferred by the outcrop pattern) is highly discordant with the fabrics developed in the Ox Mountains Succession. Furthermore no increase in strain has been observed within the Ox Mountains Succession rocks approaching the contact. These data strongly suggest that the original contact between the Ox Mountains Succession and the Raheen Barr Succession is post metamorphic with respect to the Ox Mountains Succession. This is consistent with the data observed at Sheeans (area 4, section 4.6).

4.5.9 Late structures

D3 and D4 structures are substantially reorientated by later deformation in two areas. At Attymass the S3 fabric and contemporaneous fabrics within the granodiorite, which have a northeast–southwest orientation at (G 3050 1150) are rotated to the west towards Attymass (G 2950 1150) where they have a northwest–southeast orientation. A similar effect is observed on Benalta where the S3 fabric swings from a predominently northeast–southwest attitude at (G 3900 1700) to a northwest–southeast attitude at (G 3750 1650). No minor structures have been observed associated with these large scale (2km wavelength) folds, which formed very late in the tectonic history of the inlier. It is unclear what the cause of this late deformation was. The present author notes however that the orientation of these structures is inconsistent with Taylor's correlation of these structures with regional D4.

4.6 AREA 4 : SHANVOLLEY

Metasediments of the Lismoran, Ummoon and Leckee Quartzitic Formations, deformed and metamorphosed under greenschist facies conditions, occur in a well exposed area (8km × 3km), west of the Knockaskibole fault, which separates the metasediments from the Ox Mountains Granodiorite between Sallagher (G 1616 9670) and Greenans (G 1680 0013), (Figure 4.13). Further to the northeast the fault swings northeast and separates a large (3km × 3km) wedge of granodiorite (the structure of which is discussed in section 4.9) from the remainder of the pluton. Thus between Greenans and Largan the metasediments of area 4 have an original largely unmodified contact with the Ox Mountains Granodiorite. Carboniferous rocks onlap to the north of the area, while rocks of the Raheen Barr Succession (Long and Max, 1977) are in tectonic contact with the Ox mountains Succession at Sheeans in the west. Devonian strata and the rocks of Tullycommons (area 5) are in fault contact with the Shanvolley area and form the southwest and southern boundaries respectively.

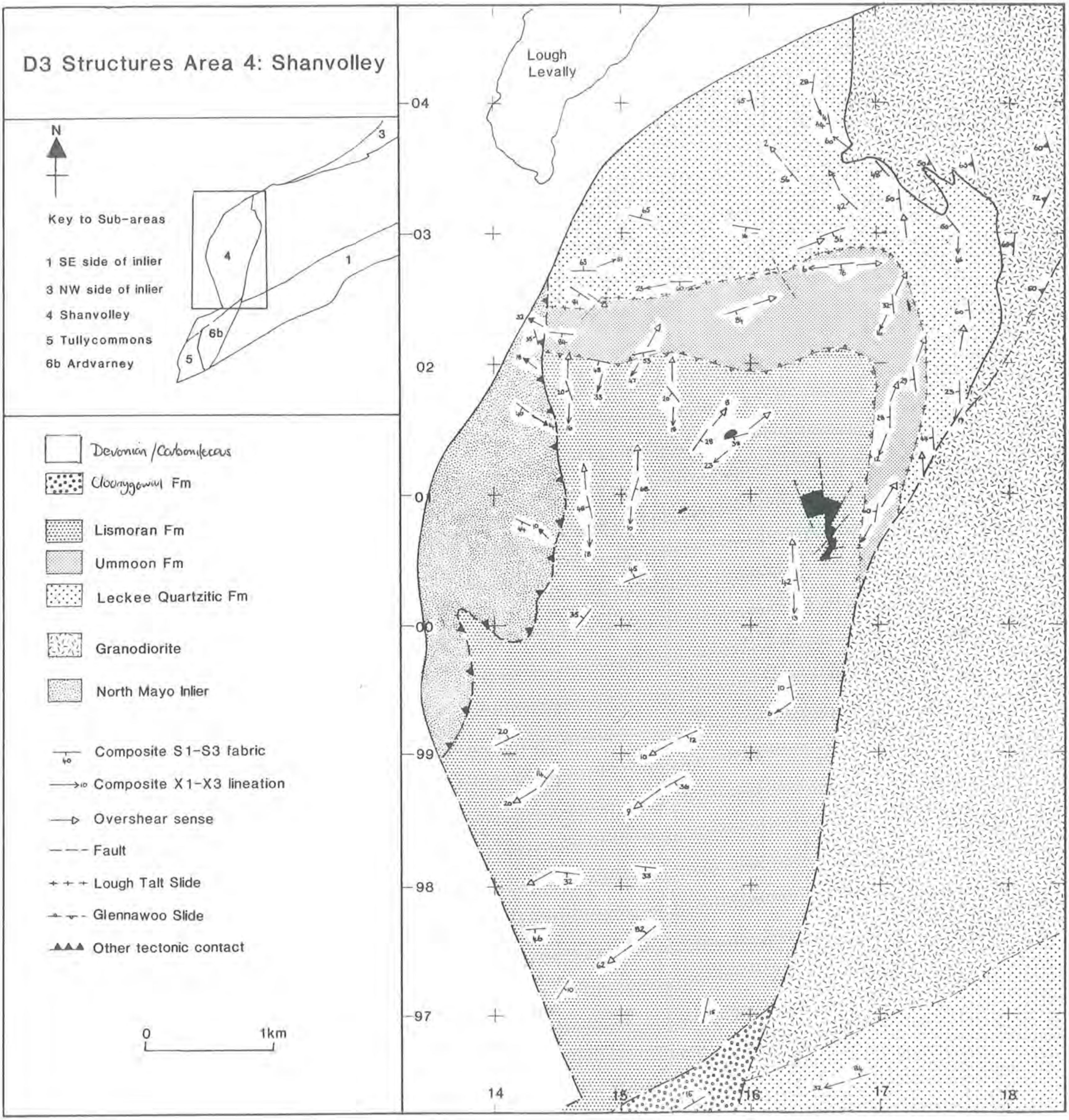
4.6.1. Ss.

Lithological banding can be distinguished throughout the area, but way up criteria have not been observed. The stratigraphy can be correlated with the remainder of the inlier and the stratigraphic sequence, with the Leckee Quartzitic Formation structurally underlying the Ummoon Formation suggests that the rocks are right way up. Lithological banding and contacts now lie parallel to S3 and are generally shallowly dipping, (<45°).

4.6.2 D1/D2

Neither quartz segregations, nor a tectonic fabric related to D1 have been observed in the Shanvolley area. Identification of the first fabric seen in the area as S2, is therefore based on correlation with the regional D2 deformation and

FIG 4.13



D3 Structures Area 4: Shanvolley

not on evidence from the Shanvolley area itself. S2 has only been observed in the hinges of later F3 folds, where it is a penetrative muscovite-chlorite-quartz fabric, orientated parallel to lithological banding. Due to its transposition into the intense S3 fabric that is developed throughout the area, neither the vergence of S2 on bedding nor F2 folds have been observed; the pre-D3 geometry can therefore not be determined. Feldspar porphyroblasts are developed pre-D3, but it is not clear whether these represent MS2 or MP2 crystallization.

4.6.3 D3

In common with the Ox Mountains Succession east of the Knockaskible fault the Shanvolley area has experienced intense deformation during D3, which is represented by isoclinal F3 folding and an S3 axial planar fabric. This is itself modified by shearbands (SC3) that are developed throughout the area.

S3 is everywhere a penetrative fabric consisting of an alignment of muscovite, chlorite and quartz. Quartz grains have a strong preferred dimensional orientation accompanied by extensive sub-grain growth and grain size reduction, indicative of high strain. F3 folds are most clearly observed west of Largan (G 1650 0235) in the northeast of the area, where their axial planes are steeply inclined to the south and their axes plunge gently to the east-northeast or west-southwest (Figure 4.14). The vergence of these structures is consistently to the north-northwest and towards the Lough Talt Slide: a similar relationship to that observed in area 1 (the southeast side of the inlier).

Figure 4.14 illustrates that the orientation of S3 varies throughout the area. The east-west strike at (G 1600 0300) in the northwest continues to Farbreiga (G 1692 0296), where it turns sharply to the southeast towards Greenans (G 1680 0013) after which it gradually turns towards the south-southwest. Initial examination of the map pattern and orientation of the stratigraphic boundaries in the northern half of the area suggests that this variable orientation of the

Figure 4.14 Stereographic projection of structural data : Area 4 (Shanvolley)

(a) S3 : mean orientation of foliation plane 119/23, (pole, 67/029), n=375

(b) $\beta 3$: mean orientation of lineation 13/208, n=275

(c) S4 : mean orientation of foliation plane 225/75 (pole, 15/135), n=92

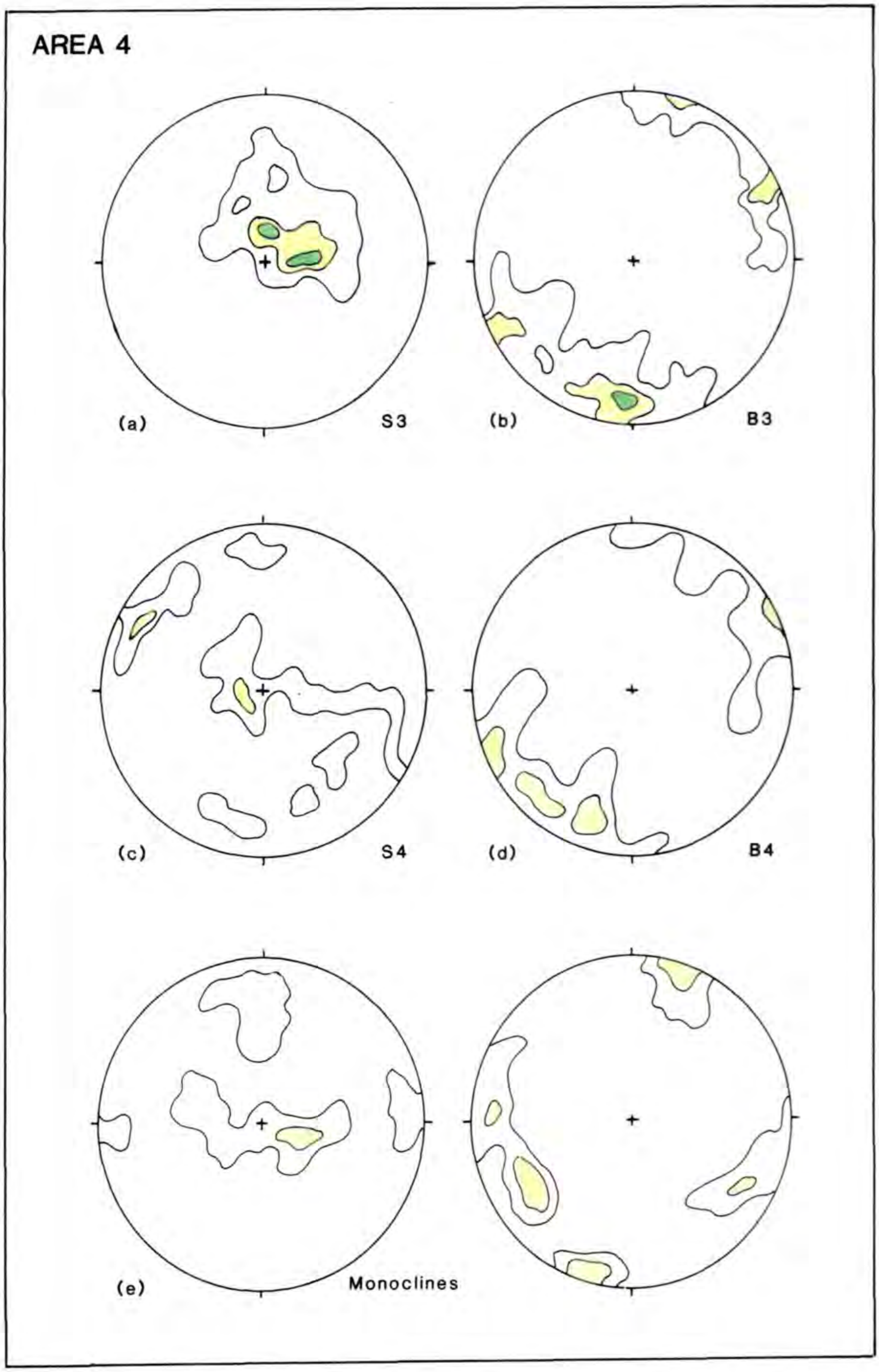
(d) $\beta 4$: mean orientation of fold axis 15/226, n=77

(e) S5 : mean orientation of axial plane 124/07 (pole, 83/034), n=54

(f) F5 : mean orientation of fold axis 01/243, n=50

Note–See Figure 4.5 for key to contour intervals.

FIG 4.14



stratigraphic contacts and S3 fabric is due to re-orientation by later folding, as the axes of F4 folds generally trend to the north-northeast (Figure 4.14).

The L3 stretching lineation does not retain a constant geometrical relationship with, nor does it show a change in orientation with the progressive reorientation of S3, but consistently trends to the north or northeast, with kinematic indicators that consistently indicate north directed overshear, (Figure 4.13). The present geometry of D3 cannot therefore simply be explained by re-orientation resulting from later folding. The data are consistent with dextral movement on the moderately west dipping Knockaskibole Fault. Matching of stratigraphy suggests that the Shanvolley area was displaced approximately 8km to the northeast relative to the remainder of the inlier to the east. The data also suggest that the fault was active late during D3, when it exerted an important influence on the structure of the area. Thus during the early stages of D3, with the Shanvolley area located south of its present position an early S3 fabric was developed axial planar to northwest verging F3 folds, similar to those seen in area 1 (the southeast side of the Ox Mountains Granodiorite). This fabric subsequently intensified to produce the tectonic slides and was further modified by shear band development as the area was translated to the north by dextral movement on the Knockaskibole fault (section 4.6.5).

Two narrow zones of increased strain named the Lough Talt Slide and Glennawoo Slide, after their type localities along the southeast side of the inlier (Taylor, 1969), occur within the Shanvolley area. The structure and kinematics of these slides is discussed in detail in section 4.3.5 (area 1), which includes the type localities, where it is shown that in the Shanvolley area the Lough Talt Slide is marked by a 20m wide zone of intense D3 strain, located at the boundary between the Leckee Quartzitic Formation and the Ummoon Formation. It is well exposed in the high ground in the north of the area at Farbreiga where quartzites of the Leckee Quartzitic Formation show mylonitic textures, totally obliterating evidence of earlier deformation. It is therefore not possible to determine whether the structure has had an earlier tectonic history. The Lough Talt Slide in this area is everywhere orientated parallel to lithological banding, it is therefore not possible to determine whether the structure has cut out any stratigraphy. Similar relationships pertain for the Glennawoo Slide, seperating the Ummoon and Lismoran Formations, in this area.

4.6.4 Ox Mountains Succession – Granodiorite contact

The Shanvolley area provides a unique situation in the inlier of a shallowly dipping contact between the metasediments of the Ox Mountains Succession and the Ox Mountains Granodiorite. Although the Knockaskibole Fault intervenes between the metasediments and the Ox Mountains Granodiorite along most of the eastern boundary of the area at (G 1680 0013) the fault swings northeast into the granodiorite and leaves the original gently inclined sheeted roof contact to the pluton isolated to the west. Moderately west dipping Sc fabrics indicate that the contact experienced dextral deformation during the later stages of D3. The Knockaskibole fault and associated splays isolate this contact area from that to the east where northeast–southwest sinistral transcurrent deformation is prevelant. Models for the tectonic development of this area are discussed in chapter 5.

4.6.5 The Knockaskibole Fault

The Knockaskibole Fault (Long and Max, 1977) located between Coarsepark (M1360 9194) and Lough Conn (G 1900 0560), was originally identified by Currall (1963) who described it as a major thrust. Currall identified two major thrust planes along which southeast directed thrusting occurred; the Sole Thrust and the structurally higher Conloon Thrust. He considered that these structures follow an identical course from Coarsepark to Greenans at which point they diverge.

The Sole Thrust follows a northeast trajectory to the shore of Lough Conn (G 1900 0560), whereas the Conloon Thrust swings to the northwest disappearing beneath the Carboniferous cover between Lough Conn and Lough Levally. These two structures therefore isolate a wedge shaped area of granodiorite in which further thrusts characterized by highly crushed and mylonitic rock have been identified (Currall, 1963). Long and Max (1977) noted that the Knockaskibole fault (Sole Thrust of Currall, 1963) had a dextral displacement of approximately 10km and that it was likely to represent a splay of the Clew Bay Fault (see chapter 6).

The present study indicates that the Knockaskibole fault has experienced a protracted history of movement. Ductile dextral fabrics with a mean orientation of 193/58W deform a wedge shaped area of the Ox Mountains Granodiorite between Greenans and Lough Conn, and overprint fabrics that are related to the earlier sinistral shear that penetratively deforms the majority of the inlier. This kinematic history is identical to that experienced by the metasediments of the Shanvolley area, which also show dextral shear fabrics. This shear sense is consistent with the fabrics having formed as a response to the dextral movement that displaced the Lough Talt and Glennawoo Slides approximately 10km northeastwards from their position in area 1a. Discrete low angle faults that dip moderately to the northwest, and are characterized by narrow (up to 1m thick) zones of brecciation and cataclasis, overprint this earlier dextral deformation and are considered to represent the final stages of movement on the structure. The southwest end of the Knockaskibole Fault also preserves evidence for the late brittle history of the structure. At (M 1452 9320) west of Slievenagark Lough, a narrow exposure gap (minimum 50m) separates the Leckee Quartzitic formation from the Cloonygowan Formation to the west. Fault related deformation here is restricted to minor brecciation in both formations. No evidence of an earlier ductile history is preserved. This suggests that at the present exposure level the structure is simply expressed as a brittle fault.

4.6.6 D4

Mesoscopic F4 folds that locally reorientate S3 occur throughout the area (Figure 4.14). They are open symmetrical folds, that often have a conjugate geometry but do not have an axial planar cleavage. Stereograms (Figure 4.14) illustrate that the F4 fold axes display a similar range of orientations to the L3 lineations. This is supported by field observations, where F4 fold axes normally share a common orientation with the L3 lineation. F4, axial planes intersect S3 at a high angle (generally >20°), suggesting that D4 represents vertical shortening or strike-perpendicular extension. Examination of thin sections confirms that in common with other areas in the inlier, D4 occurred at low metamorphic grade: muscovite and chlorite crystals are bent, while quartz grains are strained and show undulose extinction and preferred dimensional and crystallographic orientation parallel to the axial planes of F4 folds.

4.6.7 Later deformation : The contact with the Raheen Barr Succession

At Sheeans, in the northwest of the area the metasediments of Shanvolley are in thrust contact (The Sheeans Thrust) with the Raheen Barr Succession of North Mayo. The thrust contact dips gently 5-15° northwest and separates the rocks of the Ox Mountains Succession from limestones, metadolerites and grits of the structurally overlying Raheen Barr Succession. The penetrative fabric in the Raheen Barr Succession is moderately inclined to the west or southwest, and the associated west northwest–northwest orientated stretching lineation, is orientated at a high angle to the south dipping fabrics developed in the Ox Mountains Succession in the footwall. These footwall rocks show a progressive increase in brittle deformation, fracturing and quartz veining towards the contact, but do not exhibit any increase in ductile strain. The northwesterly dip of this structure suggests that it is a relatively late thrust to the southeast. This is consistent with the northwest–southeast X lineation in the hanging wall representing the original movement vector.

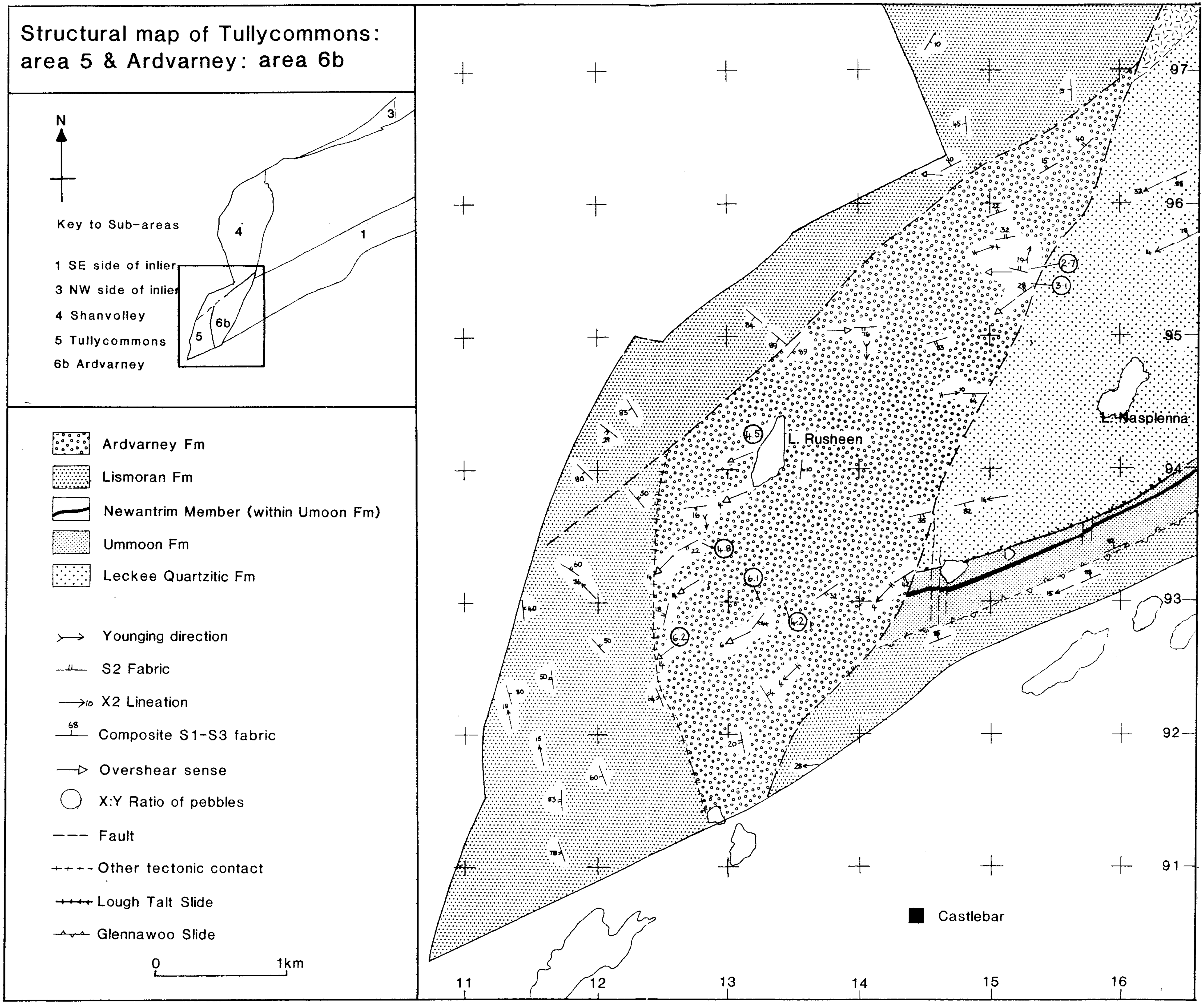
A number of small scale monoclinal folds with maximum amplitudes of 1m, whose steeply inclined axial planes are often the site of late faults with small displacements occur within the Shanvolley area. These structures have a variable trend (Figure 4.14) and consistently cut across F4 and earlier structures. It is not clear if these late folds are related to movement of the Sheeans Thrust, late movement on the Knockaskibole fault, or are due to some other cause, however their sporadic development and small amplitude (<0.5m) suggests that they do not represent significant movement.

4.7 AREA 5 : TULLYCOMMONS

These greenschist facies semi-pelites, pelites and psammites of the Ox Mountains Succession occur in a small (6km × 3km), very poorly exposed (<1%) area at Tullycommons (M 1200 9150) west of Ardvarney (area 6b), (Figure 4.15). The eastern contact between the rocks of this area and those of the Cloonygowan Formation to the east is not exposed. Carboniferous rocks onlap to the south, while a down-faulted block of Devonian strata forms the western boundary of the area. The contact with the rocks of the Raheen Barr Succession (Long and Max, 1977) outcropping to the west of the area, is not exposed.

Although the Tullycommons area exhibits evidence of a complex polyphase deformation history, it is characterised by deformation that is considerably less intense than that experienced by the Shanvolley area immediately to the north (area 3), and the remainder of the Ox Mountains Succession. The structure of the area is dominated by a generally moderately inclined penetrative fabric (S2) which is orientated parallel to lithological banding and is crenulated and folded by very rare F3 folds. F3 structures are succeeded by a later series of open

FIG 4.15



Structural map of Tullycommons: area 5 & Ardvarney: area 6b folds whose axial planes intersect lithological banding at a high angle, and have a spaced fabric developed parallel to their axial planes. The area lacks the mylonitic deformation, shear bands, and common F3 folds that characterize the remainder of rocks of the Ox Mountains Succession (excluding the Cloonygowan Formation of area 1b), which are dominated by intense transcurrent deformation.

4.7.1 Ss

Lithological banding can be distinguished widely throughout the area, but way up criteria have not been observed in this part of the inlier.

4.7.2 D1/D2

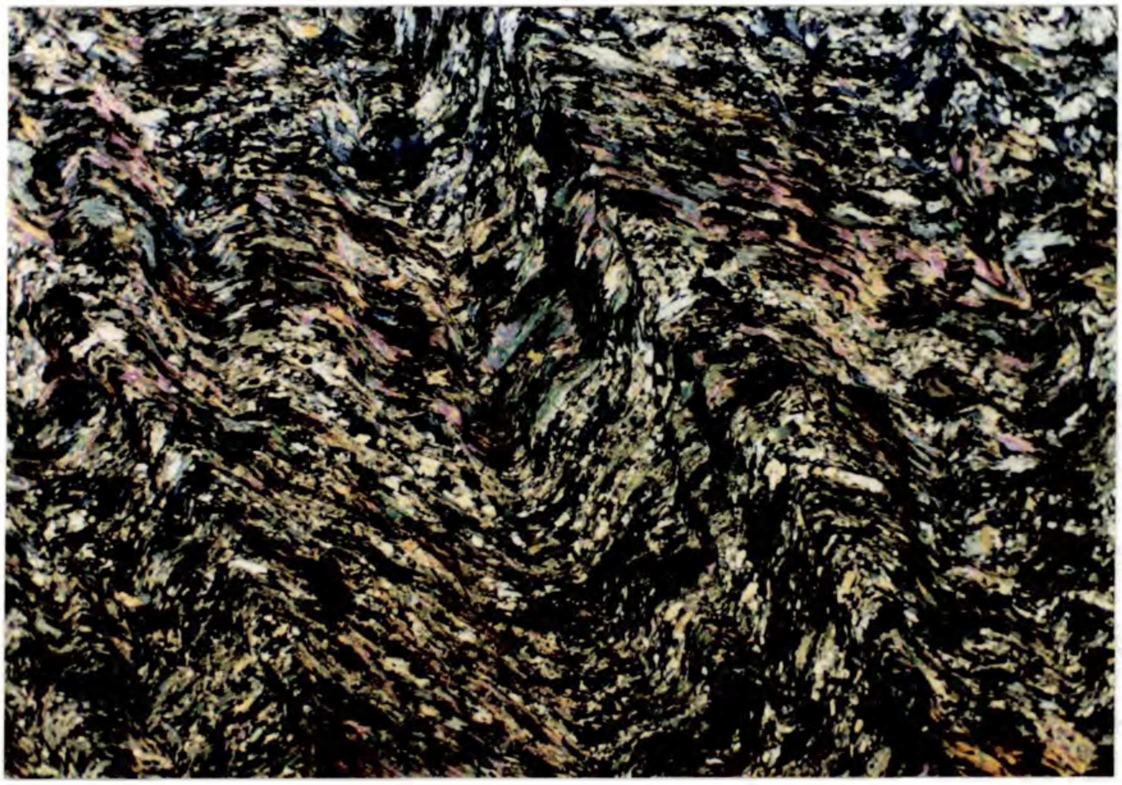
Comparison with better exposed parts of the inlier suggests that the penetrative fabric observed in the area is a correlative of the main penetrative fabric recognised throughout the remainder of the inlier (the regional D2 event) which is usually a composite S2/S1/Ss fabric in which the two tectonic fabrics cannot usually be distinguished from lithological banding. At Tullycommons, S2 is a generally north-south trending fabric that is steeply inclined (>45°), either to the west or east (Figure 4.11), and has a mean orientation 345/59E. In thin section, S2 is represented by an alignment of muscovite and chlorite (MS2). Quartz crystals which separate the bands of phyllosilicates show post-tectonic recrystallization, and are equidimensional with 120° triple points (Plate 4.28). No intensification of the S2 fabric was observed close to the unexposed contact with the Cloonygowan Formation of sub-area 6b.

4.7.3 D3

F3 folds are exceptionally rare within the area, and have only been observed at three localities. At (M 1230 9186) in the south of the area, upright, close F3 structures (Plate 4.29) fold quartz segregations that are parallel to the earlier

Plate 4.28 Photomicrograph illustrating the S2 fabric defined by a preferred dimentsional orientation of quartz, muscovite and chlorite, deformed by D4 crenulations, Lismoran formation, Tullycommons (M 1172 9156).

Plate 4.29 A rare isolated example of upright, close F3 folds with a weak discrete axial planar fabric, Lismoran Formation, close to the contact with the Cloonygowan Formation of area 6b, Tullycommons (M 1230 9186).



Field of view 14mm

S2 penetrative fabric. F3 folds verge towards the southwest, and their axes plunge gently to the southeast. A weak discretely spaced fabric (S3) developed parallel to the axial planes of the folds can be distinguished in the fold hinges, where it intersects S2 at a high angle. In thin section muscovite and chlorite has crystallized parallel to S3. Similar relationships can be observed at (M 1230 9380) and (M 12240 9330), but vergence data could not be obtained from these small outcrops, the geometry of D3 in area 5 is therefore very poorly constrained.

No intensification of the D3 deformation has been recorded close to the unexposed contact with the adjacent rocks of the Cloonygowan formation. Although the S3 fabric is sporadically developed adjacent to the contact (Plate 4.29) it has a highly variable orientation which is considered to be inconsistent with the presence of a shear zone or tectonic contact here.

4.7.4 D4

A shallowly inclined (generally <25°) (Figure 4.16) discrete fabric with an average spacing of 1mm, is widely developed throughout the area, but is most conspicuous where it affects pelitic lithologies. It is most clearly observed at (M 1772 9156) where it is parallel to the gently inclined axial planes of north north east trending open F4 folds that gently fold bedding and the dominant S2 fabric. In common with the remainder of the inlier no recrystallization is associated with this fabric, MS3 muscovite and chlorite crystals are bent, and quartz shows undulose extinction, indicating that deformation occurred at very low metamorphic grade.

The area is therefore considered by the present author to have experienced a tectonic history that is distinct from the rest of the Ox Mountains, which is discussed in the following sections. In brief, the present author considers that the area must have been spatially remote or isolated from the Ox Mountains shear zone during the later stages of D3, when the rest of the inlier experienced

Figure 4.16 Stereographic projection of structural data : Area 5 (Tullycommons)

(a) S2 : mean orientation of foliation plane 314/44, (pole, 46/224), n=33

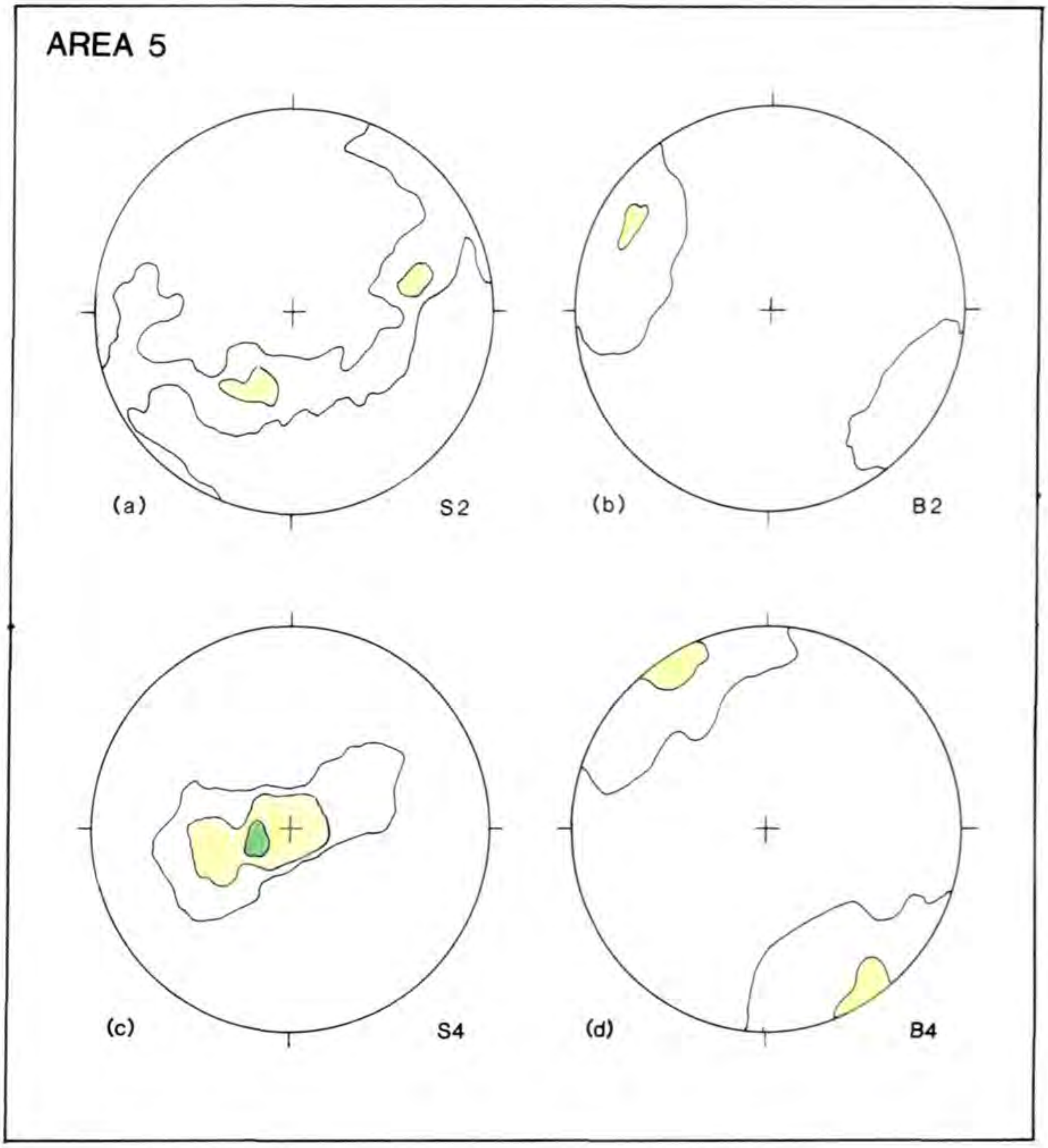
(b) β2 : mean orientation of lineation 01/155, n=18

(c) S3 : mean orientation of foliation plane 337/12 (pole, 78/267), n=15

(d) β3 : mean orientation of lineation 01/155, n=18

See Figure 4.5 for key to contour intervals.

FIG 4.16



intense transcurrent deformation. This implies that the east-west trending fault (The Lough Ben Fault) that separates Tullycommons from Shanvolley (area 4) to the north, and the Knockaskibole fault which separates it from the remainder of the Ox Mountains Succession to the east have significant displacement and have exerted considerable control on the tectonic development of the inlier. This is discussed fully in chapter 5.

4.8 AREA 6 : THE CLOONYGOWAN FORMATION

Rocks of the Cloonygowan Formation which were deformed and metamorphosed under low-greenschist facies conditions are exposed in three separate areas:

1. Cloonygowan (sub–area 6a), situated on the southeast side of the inlier.
2. Ardvarney (sub–area 6b), 15km southeast along strike.
3. Westport (sub–area 6c), 12km southeast of the Ox Mountains inlier.

All the sub-areas have experienced a polyphase deformation history under greenschist facies conditions (Figure 4.4), but the geometry and relative intensity of the deformation phases differs between sub–areas 6a and 6c, and 6b. This is considered by the present author to reflect their different position within the shear-zone. Therefore the deformation chronology and nature of the contacts with the surrounding areas are separately examined for each of the three sub-areas. These data are summarized in figure 4.17 and the evidence for the correlation of the structural history of each of the sub-areas is reviewed.

4.8.1 Cloonygowan (sub–area 6a)

Low grade greywackes of the Cloonygowan Formation are exposed in a small (5km × 1km) lozenge shaped area at Cloonygowan (G 3300 0230) on the southeast

	CLOONYGOWAN (area 6a)	ARDVARNEY (area 6b)	WESTPORT (area 6c)
D1	S1 grain shape alignment fabric	Weak bedding parallel alignment of musc-chl	
D2	S2 penetrative fabric inclined at a small angle to bedding No shear bands observed	S2 penetrative fabric inclined at a small angle to bedding. F2 folds close-isoclinal, hinges now isolated by high strains. S2 intensifies westwards towards the contact with the Ox.Mts.Succn. Shear bands develop, indicating SW overshear.	S1 and S2 not distinguished. S2 penetrative fabric, inclined at a small angle to bedding. S2 fabric consistently dips NW. Local younging reversals indicate the presence of mesoscopic F2 folds. No shear bands observed.
D3	S3 discretely spaced pressure solution fabric intersects S2 at a high angle Sinistral shear bands developed adjacent to contact with Ox Mts. Succn.	Rare steeply inclined spaced pressure soln. fabric crenulates S2.	S3 spaced pressure solution fabric intersects S2 at a high angle. Vergence on S2 consistently NW. F3 tight-isoclinal folds, max amplitude 1m consistently verge north. Sinistral shear bands
D4	F4 open folds, axes plunge NE, axial planes inclined gently NW or SE, no axial planar fabric.	Open folds common. Axes plunge gently NE-SW, parallel to X2 Shallowly inclined axial planes.	F4 open folds common. Axes plunge gently NE-SW. Axial planes shallowly inclined at a high angle to S3.
D5	Kink bands with subvertical axes, offsets consistently dextral.	Later deformation uncommon, occasional dextral kinkbands. Entire area reorientated by later faulting.	

FIG 4.17

side of the inlier. An exposure gap of 50m separates this area from the Ox Mountains Succession to the northwest, while Carboniferous rocks onlap to the southwest and form the southeast margin of the inlier.

4.8.1.1 Ss : Lithological banding, marked by substantial variations in grain size, can be recognized over most of the Cloonygowan area. Graded bedding is common (Plate 4.30, and therefore the way up of strata can normally be easily determined. Lithological banding has been greatly re-orientated by later deformation. In particular, F3 folding has steepened lithological banding so that it is now generally inclined at > 45 °. Way up criteria observed suggest that the strata are frequently overturned by earlier mesoscopic F2 folds.

4.8.1.2 D1 : S1 is clearly observed at one locality in sub–area 6a of the Cloonygowan Formation. At (G 3300 0190) S1 is expressed as a grain shape alignment fabric which is inclined at a small angle to bedding and downward facing. No minor folds associated with this fabric have been observed. Similar relationships are seen in other parts of the area but, generally, D2 and D3 deformation transposes S1. There is therefore insufficient information to deduce the geometry of D1. The identification of S1 within the Cloonygowan Formation is however of critical importance as it enables a detailed correlation to be made between structural chronology of the rocks of the Cloonygowan Formation and those of the Ox Mountains Succession which is discussed in detail in section 4.10.

4.8.1.3 D2 : The regional S2 fabric is developed throughout the Cloonygowan area where it is expressed as a penetrative fabric inclined at a small angle to lithological banding. S2 consists of a preferred dimensional and crystallographic orientation of fine-grained (0.2 × 0.04mm) quartz, muscovite and chlorite. Detrital quartz and feldspar grains are strongly flattened into the plane of S2 and the X-direction of the detrital grains lies parallel to the X3 stretching lineation which is parallel to the intersection of S2 and S3. S2 is axial planar to a number of rare northeast-southwest trending, recumbent, mesoscopic, tight-isoclinal folds that commonly face to the southeast and produce numerous younging reversals, (Phillips et al. 1975). D2 structures do not show consistent vergence relationships, which suggests that the sub–area occupies a D2 hinge zone. Large scale F2 folds closures could not be mapped out within the limited area of exposure.

Phillips et. al. (1975) state that this D2 fabric intensifies towards the northwest and the unexposed contact with the Ox Mountains Succession, which they regard as a tectonic contact of D2 age. No evidence, such as a decrease in the spacing of S2 or a decrease in the angle between S2 and Ss, was observed by the present author to confirm that D2 strain increases towards the contact in either the Cloonygowan Formation or the Ox Mountains Succession.

4.8.1.4 D3 : S3 can be identified easily throughout sub–area 6a where it is a steeply inclined pressure-solution cleavage (Plates 4.30–4.31) with a mean orientation of 059/85 SE, (Figure 4.18). The strong X3 stretching lineation lies sub–parallel to the S2/S3 cleavage intersection and plunges gently to the northeast, with a mean orientation of 06/061.

The spacing of S3 decreases towards the northwest, reflecting the increase in D3 strain as the contact with the Ox Mountains Succession is approached (Figure 4.19). This is supported by measurement of the shortening of post-D2 quartz veins which show an increased shortening from 10% 700m from the contact to 90% at the exposures closest to the contact, which is itself unexposed. The increase in D3 strain is also reflected by the development of sinistral shear bands in the last 50m of exposure (Figure 4.19). The C planes of the shear bands, which bend but do not recrystallize S3 muscovite and chlorite, make an angle of up to 30° with S3 and have a sub-vertical intersection with S3 which is perpendicular to the well developed X3 lineation, (Plates 4.32–4.33). Upright, close to tight F3 folds, which refold S2, are widely developed throughout the area. Their amplitude : wavelength ratio increases towards the northwest. F3 vergence data indicate

Plate 4.30 Steeply inclined S3 pressure solution fabric crenulating S2 which is orientated at a small angle to lithological banding, Cloonygowan Formation, Cloonygowan (G 3300 0184).

Plate 4.31 Photomicrograph of specimen illustrated in Plate 4.30, showing the spaced character of S3 and its crenulation of the earlier penetrative S2 foliation.

Field of view 14mm

Figure 4.18 Stereographic projection of structural data : Area 1a (Cloonygowan)

(a) SS/S1/S2 : mean orientation of foliation plane 254/80, (pole, 10/164), n=20

(b) Kink Bands : orientation of axial plane 164/84 (pole,06/074), n=99

(c) S3 : mean orientation of foliation plane 059/85 (pole, 05/329), n=148

(d) $\beta3$: mean orientation of lineation 09/061, n=73

(e) S4 : mean orientation of foliation plane 024/42, (pole, 48/294), n=22

(f) $\beta4$: mean orientation of lineation 27/060, n=50

See figure 4.5 for contour intervals

FIG 4.18

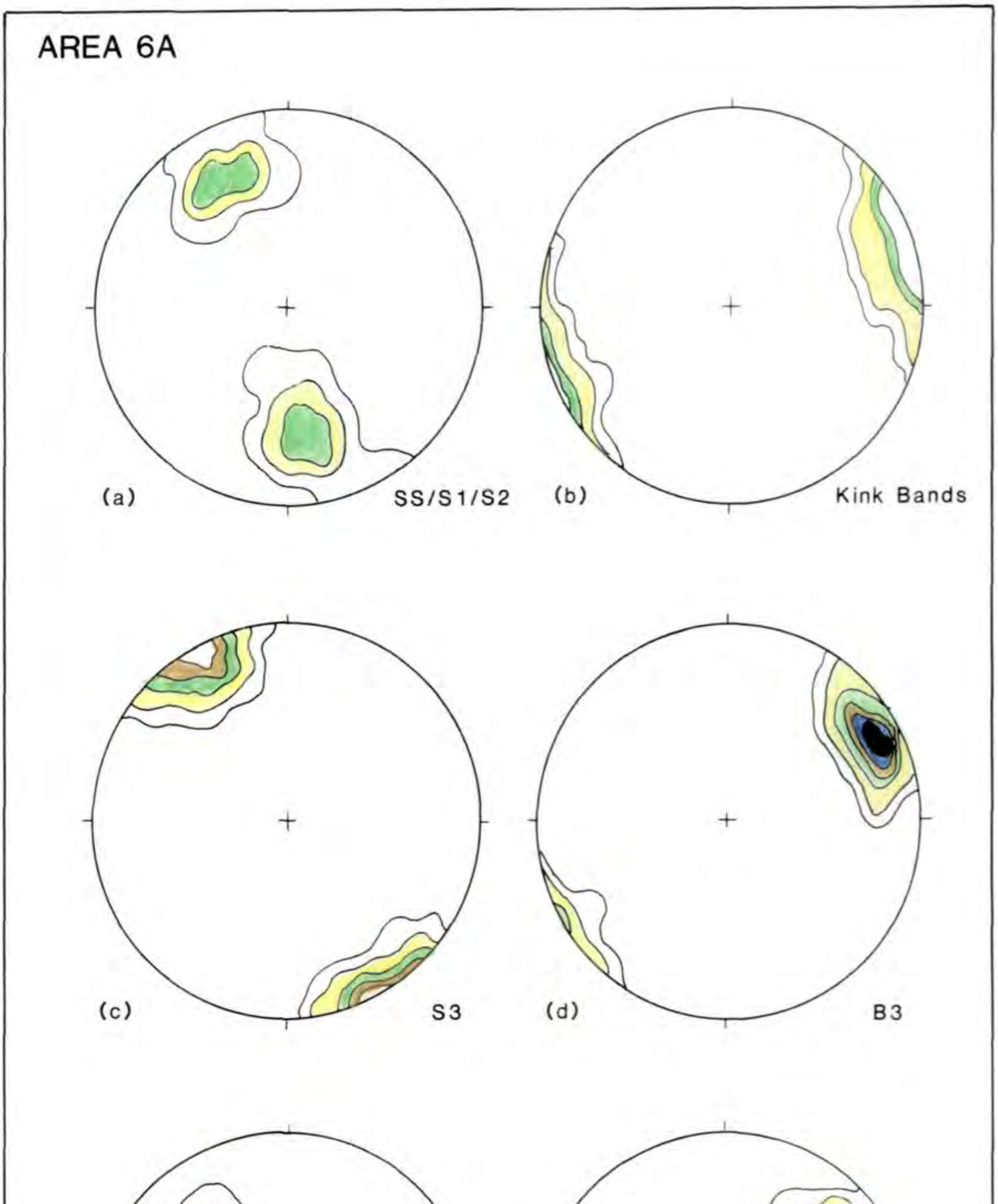
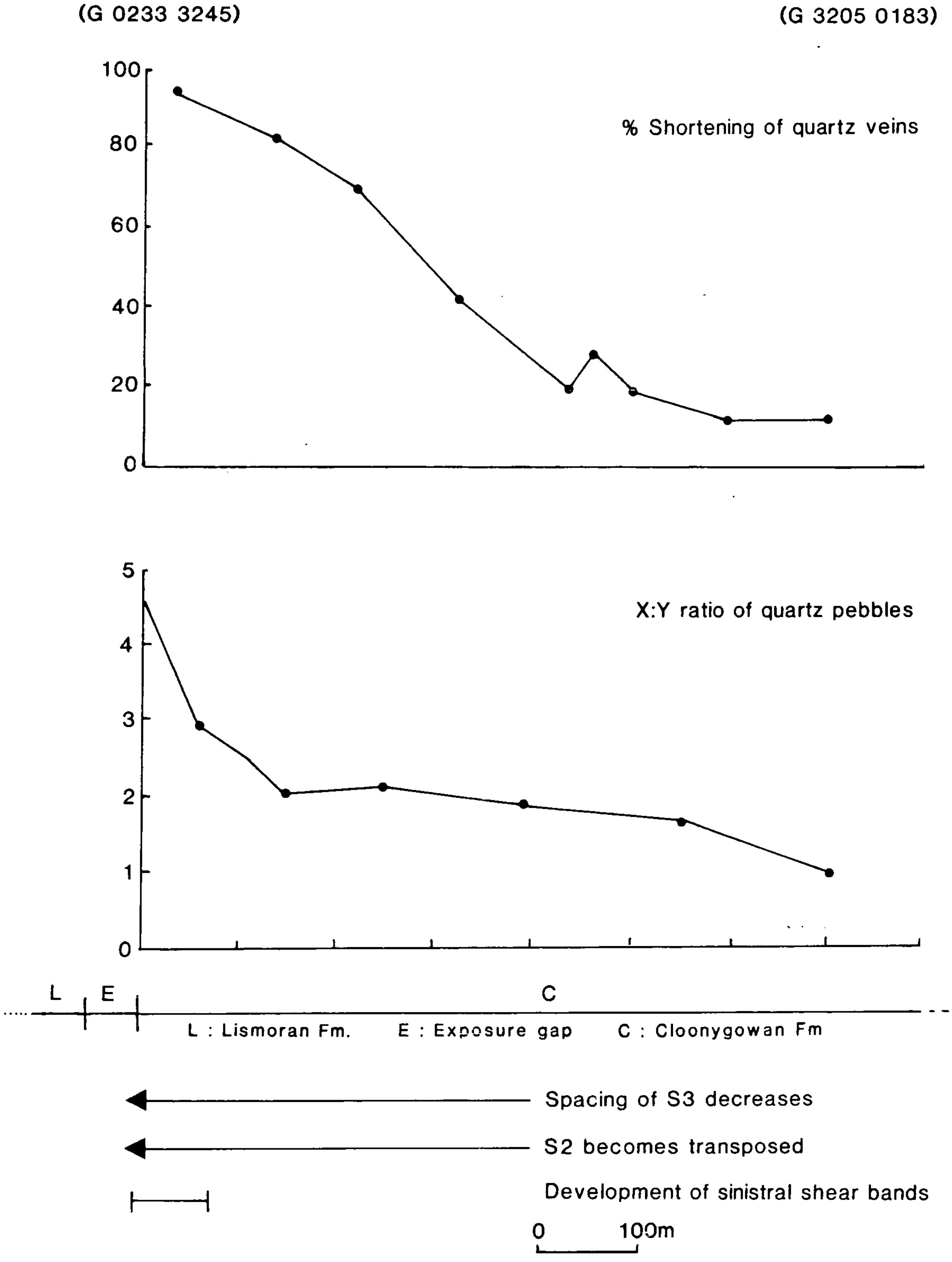


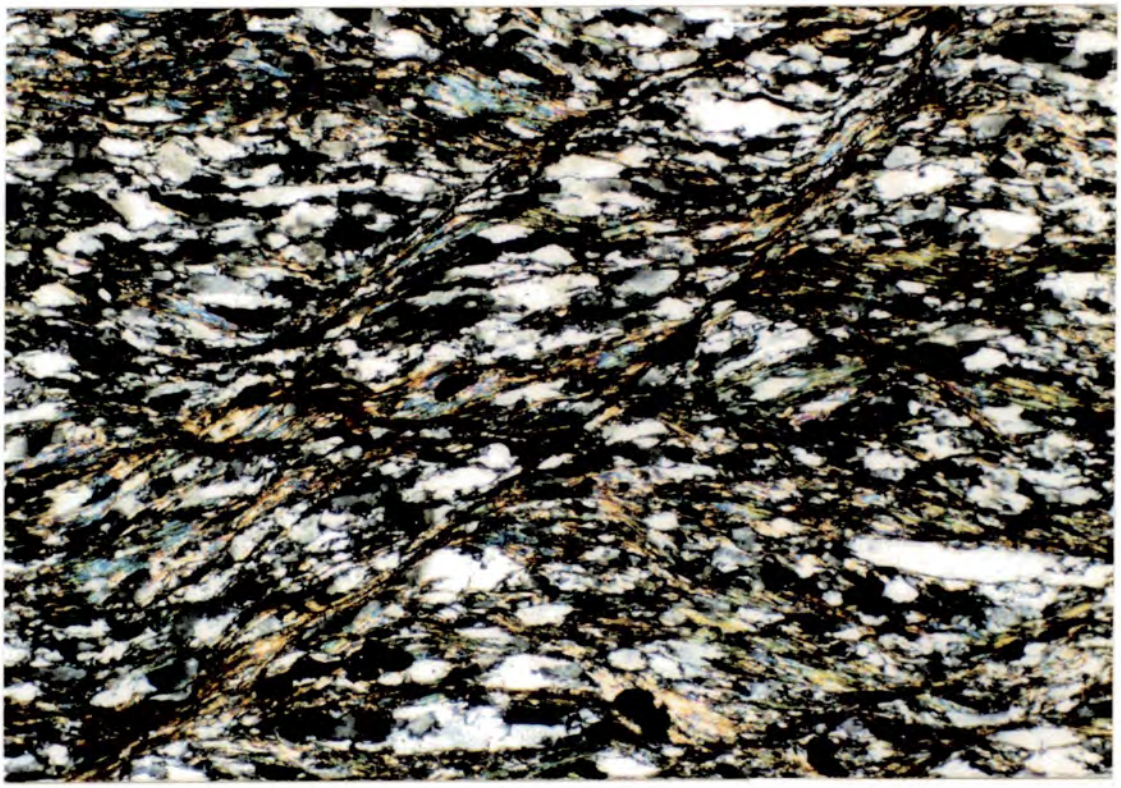
FIG 4.19



STRAIN PROFILE ACROSS THE CLOONYGOWAN FM AT CLOONYGOWAN

Plate 4.32 Sinistral shear bands, viewed looking southeast along a gently inclined suface, Cloonygowan Formation, Cloonygowan (G 3247 0226).

Plate 4.33 Photomicrograph of specimen illustated in Plate 4.32, showing the well–developed S3 foliation that is modified by shear band development.



Field of view 14mm

that the structure of the area is dominated by a major F3 synform whose axis is located at (G 3306 0215).

4.8.1.5 D4 : produced a set of open to close folds that frequently show a congugate geometry and are most commonly developed in the pelitic and semi-peltic lithologies. S4 axial planes fall into two groups, (Figure 4.18, Plate 4.34) which dip moderately to the northwest and less commonly to the south or southeast.

F4 axes plunge gently to the northeast with a mean orientation of 27/060 (Figure 4.18), parallel to the well-developed X3 stretching lineation. MS3 muscovite and chlorite are bent by D4, but show no recrystallization, which is consistent with their development under very low grade conditions of metamorphism. The geometry of these structures suggests that they represent vertical shortening, or strike-perpendicular extension.

4.8.1.6 Later structures : Kink bands whose axes have a mean orientation of 84/164 (Figure 4.18), intersect S3 at a high angle and are common throughout the area. 98% have a dextral displacement on S1–S3.

4.8.1.9 The Callow Fault : identified by Taylor (1968), the Callow Fault is concealed beneath a thick cover of peat, and is nowhere directly observed to fault rocks of the Cloonygowan Formation or the Ox Mountains Succession. Faulted Carboniferous rocks exposed immediately to the south of the unexposed boundary between the two formations however indicate that a fault separates the Cloonygowan Formation from the Ox Mountains Sucession to the northwest (Long, pers comm). Earlier movement on this structure is inferred from metamorphic considerations discussed in section 4.10.

4.8.2 Sub-area 6b : Ardvarney

The greywackes of the Cloonygowan Formation occur in a small (6km × 2km) lozenge shaped area at Ardvarney. Topographic relief is low and exposure

Plate 4.34 F4 fold, folding S3, Cloonygowan Formation, Cloonygowan (G 3248 0024).

consequently very poor, occuring principally in the Clydagh river (M 1510 9559). Other areas of outcrop are typically separated by up to 1km of unexposed ground. The moderately west dipping Knockaskibole fault, described in section 4.4.5, defines the eastern limit of the area, while the contact with the Ox Mountains Succession to the west is not exposed. Along its northern margin the area is in fault contact with area 4, (Shanvolley). Carboniferous rocks onlap to the south, forming the southern margin of the unit.

4.8.2.1 Ss : Lithological banding is visible over much of the Ardvarney area, where it generally dips at shallow angles (<30°). The way up of the strata can be determined where graded bedding is present, particularly in the coarser grained fractions of the Formation. These data indicate that the rocks are largely right way up in the Clydagh river section (M 1510 9559), although F2 folds locally overturn strata (M 1510 9560) (Plate 4.35) Further evidence that the strata are right way up is present at (M 1325 9400), 2km to the southwest, and at (M 1425 9645). Due to poor exposure and intense deformation it has not been possible to determine the way up in the rest of the area. However the consistently right way up section through the Clydagh River between (M 1522 9500) and (M 1425 9650), suggests that large scale folds have not affected the area.

4.8.2.2 D1 : A weak preferred dimensional orientation of quartz, muscovite and chlorite, that possibly represent MS1 crystallization, has been identified parallel to lithological banding in the fine-grained pelitic lithologies. This occurs in an area of anomalously low strain in the Clydagh River section at (M 1540 9522) where it is folded by F2 folds (Plate 4.35). No minor folds have been observed which are associated with this S1 fabric.

4.8.2.3 D2 : D2 is represented by a penetrative fabric, produced by MS2 crystallization of quartz, muscovite and chlorite. That is developed throughout the area, and lies at a small angle to bedding (Plate 4.36). The geometry of D2 is most clearly observed in the small area of fine-grained pelitic rocks in the Clydagh

Plate 4.35 Close F2 folds, with sub–horizontal axial planes, Cloonygowan Formation, Clydagh River, Ardvarney (M 1510 9559).

Plate 4.36 Photomicrograph illustrating the bedding parallel S2 fabric, Cloonygowan formation, Clydagh River, Ardvarney (M 1540 9522).

Field of view 14mm

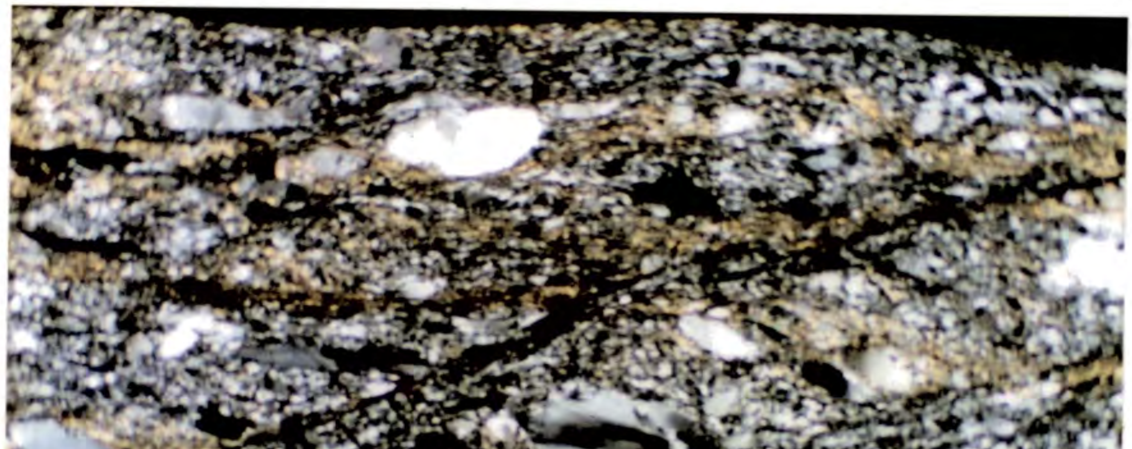
River (M 1510 9559) described above, that exposes the only examples of F2 folds recorded at Ardvarney (Plate 4.35). Ss/S1 is deformed by a series of tight - isoclinal folds (average amplitude 15cm, wavelength 6cm), with shallowly inclined axial planes. The Ss/S1 - S2 intersection lineation makes a high angle with the X2 stretching lineation indicating that the strain is low (elsewhere in the inlier F2 and F3 fold axes lie parallel to X3). At this locality F2 folds face to the southeast and verge towards the northwest. However as F2 folds have not been recorded elsewhere in the area it is not possible to determine the geometry of D2 in more detail. As discussed chapter 2 the stratigraphy of the area is generally right way up. This strongly suggests that F2 folds only have a minor influence on the structure of the area, and that larger scale F2 folds are either absent or are on such a large scale that they cannot be appreciated within the small area of available outcrop.

In the remainder of the Ardvarney area, which has been subjected to higher strain, S2 transposes the Ss/S1 fabric, which becomes a composite Ss/S1/S2 foliation into which detrital grains of quartz and feldspar are flattened. This composite fabric is modified by shear band development accompanied by grain size reduction and sub-grain growth (Plates 4.37–4.38). The composite fabric is usually gently inclined and has a mean orientation 250/10NW, (Figure 4.20) the associated X2 stretching lineation is orientated 02/062.

A D2 strain gradient, increasing towards the west and the unexposed contact with area 3, can be mapped by recording the X:Y ratios of pebbles within the coarse grained lithologies that have been flattened into the S2 fabric. X:Y ratios increase from 2.7 at (G 1505 9586) to 6.2 at (M 1255 9315) as S2 intensifies, and becomes a mylonitic fabric. Pressure shadows are developed around the quartz and feldspar clasts, indicating a rheology contrast between matrix and pebbles. The strong stretching lineation associated with this mylonitic fabric has a mean orientation of 02/062 (Figure 4.20), while kinematic indicators (mica fish,

Plate 4.37 Mylonitic S2 fabric modified by extensive shear band development, Cloonygowan Formation, Ardvarney (M 1510 9561).

Plate 4.38 Photomicrograph of specimen illustrated in Plate 4.37 showing the grain size reduction and recrystallization associated with shear band development.



Field of view 14mm

Figure 4.20 Stereographic projection of structural data : Area 6b (Ardvarney)

(a) SS/S1/S2 : mean orientation of foliation plane 250/10, (pole, 80/160), n=53

(b) $\beta2$: mean orientation of lineation 02/062, n=51

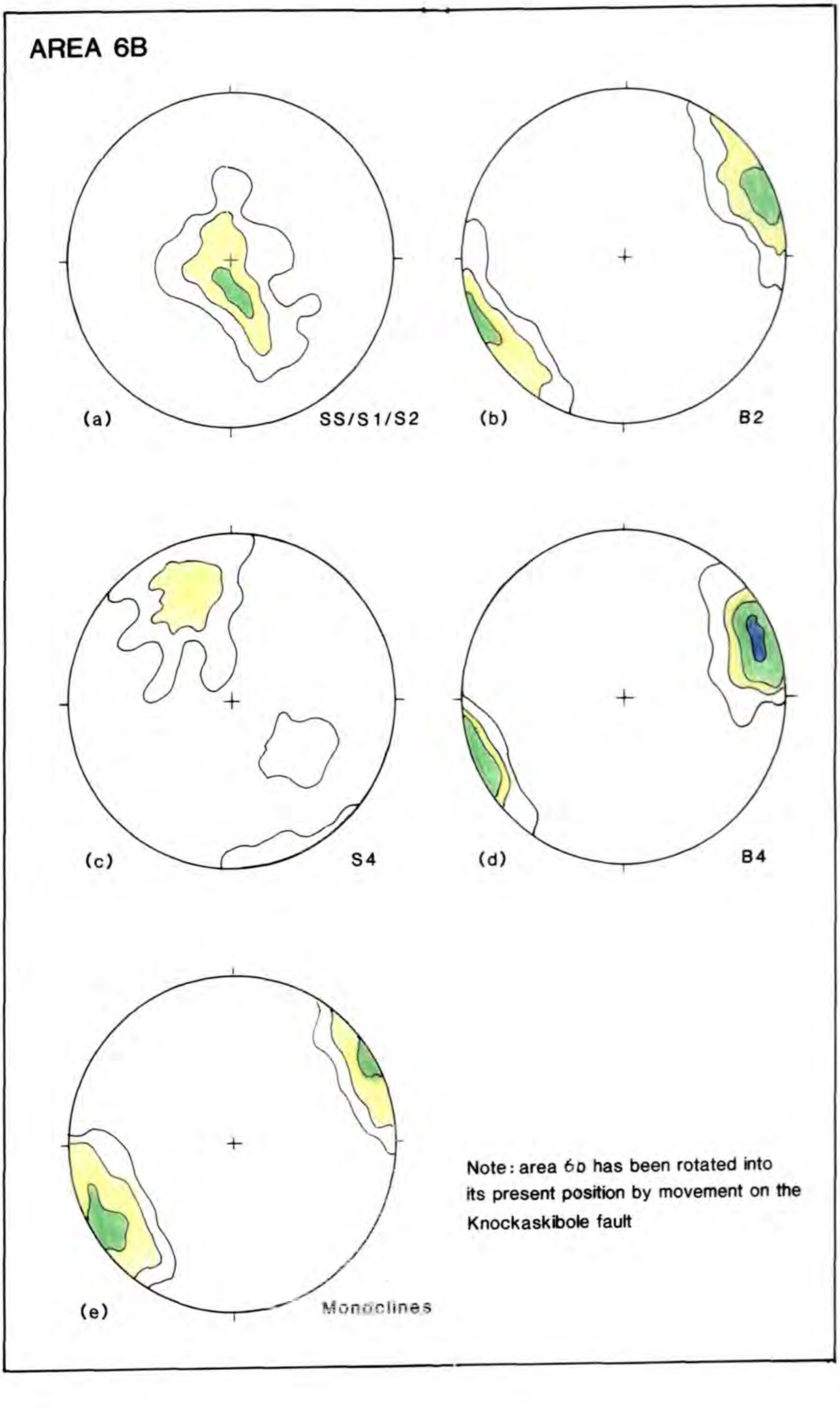
(c) S4 : mean orientation of foliation plane 056/62 (pole, 28/326), n=46

(d) $\beta4$: mean orientation of lineation 09/067, n=38

(e) Monoclinal folds : orientation of axial plane 332/82 (pole, 18/242), n=11

See figure 4.5 for key to contour intervals.

FIG 4.20



Note: area 6b has been rotated into its present position by movement on the Knockaskibole fault rotated porphyroblasts, and shear bands consistently indicate overshear towards 240 azimuth, and the contact with area 5 (Tullycommons), which is discussed fully in section 4.6.

Correlating tectonic fabrics developed in the Ardvarney area with the other areas of low-grade rocks (central Ox Mountains, area 6a, and Westport, area 6c) is problematic principally due to poor or non-exposure of the contacts between Ardvarney and adjacent areas. The correlation which is favoured by the present author is tabulated in Figure 4.17 and hinges on the correlation of the first penetrative fabric seen at Ardvarney with the regional D1 deformation. It follows from this correlation that their is no obvious correlative in the Ardvarney area for the steeply inclined S3 fabric observed at Cloonygowan and Westport, where it intersects the S2 fabric and lithological banding at a high angle. This implies that there is a significant difference in tectonic history of the areas, and that the Ardvarney area has not been subjected to the intense D3 deformation that characterizes the other two areas. Mechanisms which may account for this are discussed in chapter 5.

The alternative correlation preferred by Long (pers comm. 1988) of the penetrative fabric observed at Ardvarney with the spaced S3 fabric at central Ox Mountains is not accepted for the following reasons.

1. The fabric correlated by Long has different relationships to lithological banding. The fabric in sub–area 6b (present D2) is inclined at a small angle to lithological banding, this suggests a correlation with D2 at sub–area 6a and not D3, which intersects lithological banding at a high angle.

2. The present author does not consider that the penetrative fabric observed in sub–area 6b (D2 present chronology) is sufficiently intense to have completely transposed two earlier fabrics as required by the model of Long. This is supported by thin section examination (Plate 4.36), which shows that the S2 fabric, inclined

at a small angle to lithological banding is a closely spaced penetrative fabric, which is highly distinct in character from the spaced nature of S3 elsewhere in the inlier,

3. The S3 (present chronology) fabrics observed in sub–areas 6a and 6c are steeply inclined and have a constant orientation, which is consistent with their having formed in a shear-zone, whereas the only penetrative fabric observed in sub–area 6b is shallowly inclined and does not exceed a dip of 45°.

4.8.2.4 D3 : The Ardvarney area lacks the intense discretely spaced pressure solution fabric orientated at a high angle to bedding and S2, that characterises the rocks of sub-area 6a, where sinistral shear bands occur in the area of most intense D3 strain towards the contact with the Ox Mountains succession.

Isolated examples of a weak discrete spaced fabric that crenulates S2 and is axial planar to close, upright folds are observed at (M 1246 9184) close to the contact with the Tullycommons area (area 5) to the west. This fabric is sporadically developed and has a variable orientation. It is therefore not considered to be sufficiently intense, widely developed or regular in orientation to represent a tectonic contact of D3 age.

This striking dissimilarity between the expression of D3 within rocks of the same formation, deformed and metamorphosed at the same metamorphic grade implies that both areas must have had a different tectonic setting during D3. The tectonic model considered to explain this difference in structural history is discussed fully in chapter 5.

The relative absence of D3 deformation means that the Ardvarney area preserves critical evidence on the nature of the D2 tectonic contact that separates the Ox Mountains Succession from the Cloonygowan Formation. Subsequent movement on the Knockaskibole Fault system and related structures has rotated this contact so that it is now inclined oblique to regional strike.

4.8.2.5 Nature of the contact between the Ox Mountains succession (Tullycommons, area 5, and the Cloonygowan Formation at Ardvarney (sub–area 6b). The contact between sub–area 6b and area 5 is not exposed. The rocks of the Ox Mountains Succession and Cloonygowan formation are in closest proximity at Tullycommons (M 1246 9184) where an exposure gap of 80m separates the two units. As discussed in section 4.7.3 D2 strain increases towards this contact, indicating that it was active as a D2 structure. The sporadic development of a later spaced fabric in the vicinity of the contact (correlated with the D3 deformation of area 1a) is not considered to be sufficiently intense, widely developed or regular in orientation to represent a major tectonic contact of D3 age.

The D2 age for the contact here contrasts with the D3 age demonstrated for the contact between the Cloonygowan formation of sub– area 6a and the Ox Mountains Succession of area 1. Three explanations may account for this. The first is that it is possible that the rocks of the Cloonygowan Formation presently exposed in sub–area 6a represent a significantly higher part of the formation. If this were the case then the original contact with the rocks of the Ox Mountains Succession to the northwest would be located below the present exposure level and any original D2 contact would therefore not be apparent. The second possibility is that the intense D3 deformation that affects the Cloonygowan Formation of sub–area 6a masks an earlier D2 tectonic contact. The third possible explanation is that the rocks of the Cloonygowan Formation of the two sub areas were brought into contact with the Ox Mountains Succession at different times. This is not considered to be likely by the present author as it would imply widely differing tectonic histories for the two sub–areas.

4.8.2.6 D4 : Structures asociated with D4 are weakly developed throughout the sub–area although they are most common in the pelitic lithologies such as in the Clydagh River at (M 1510 9559). D4 is expressed by a series of small scale open to close angular folds with wavelengths generally less than 20cm whose moderately inclined axial planes intersect the well–developed S2 fabric parallel to the X2 stretching lineation. F4 fold axes have a mean orientation 09/067., (Figure 4.20). Although occasionally F4 folds form congugate sets, usually only one set is found with neutral vergence. S4 axial planes are inclined at a high angle (70-90°) to S2, which appears to control their orientation (Figure 4.20, Plate 4.39). No penetrative fabric associated with F4 folds has been observed in the Ardvarney area. These structures have clearly formed at a very low metamorphic grade. Early muscovite and chlorite grains are bent around the hinges of F4 folds but have not recrystallized. Quartz grains do, however, show evidence of MS4 syntectonic recrystallization, having a preferred crystallographic and dimensional orientation parallel to the F4 axial planes.

4.8.2.7 Later deformation : The fractured monoformal folds, and dextral shear bands that are conspicuously developed in the Cloonygowan Formation at area 6a, have not been observed in the Ardvarney area.

4.8.3 AREA 6C WESTPORT

In this section the deformation chronology established during a brief reconnaisance study of the Westport Grit Formation (Max 1989) of sub–area 6c will be examined and compared with that of the Cloonygowan formation of areas 6a and 6b in Figure 4.17.

4.8.3.1 Ss/D1/D2 : The earliest fabric observed is a spaced fabric that is attributed to composite S1–S3 and is inclined shallowly to the north and lies sub-parallel to lithological banding. Pebbles are strongly flattened into the plane of the penetrative foliation, which is developed throughout the area. In the areas examined, the strata generally young to the south, but local younging reversals indicate the presence of folds of F2 or possibly F1 age.

Plate 4.39 Open F4 folds of S2 foliation, Cloonygowan Formation, Clydagh River, Ardvarney (M 1510 9559).

4.8.3.2 D3 : D3 is expressed as a more discretely spaced (average 5mm) pressure solution cleavage that is inclined steeply to the northwest and therefore intersects S2 at a high angle. This constant northwards vergence suggests that the area is located on a single limb of an F3 fold (Plate 4.40). Sinistral shear bands which modifiy the pre–existing S3 fabric are also occasionally observed.

4.8.3.3 D4 : Shallowly plunging, mesoscopic, open F4 folds, clearly refold F3 and associated structures. They have a shallowly inclined axial plane but lack an associated axial planar cleavage and are sporadically developed throughout the area, (Plate 4.41).

The correlation of the Westport grit Formation with the Cloonygowan Formation (Chapter 3) is strongly supported by the similarity in deformation sequence observed in both areas. The D2-D4 deformation phases exhibit a similar geometrical relationship to bedding in both areas and are therefore considered to be correlatives.

4.9 STRUCTURE OF THE OX MOUNTAINS IGNEOUS COMPLEX.

In this section the deformation of the three components of the igneous complex are examined separately. Possible emplacement mechanisms are discussed in chapter 5.

4.9.1 The Ox Mountains Granodiorite

The Ox Mountains Granodiorite, which is the main component of the igneous complex is pervasively deformed by a strong penetrative fabric that is a solid–state alignment of plagioclase, biotite and alkali–feldspar (Mc Caffrey, pers comm). A sinistral S/C fabric is widely developed. This fabric has an identical orientation to the S3 fabric in the metasediments at the contacts and in rafts

Plate 4.40 Steeply inclined pressure–solution fabric (correlated with S3 in the Cloonygowan Formation) deforming an earlier (S2) bedding parallel fabric, Westport Grit Formation, Westport (M 0200 8450).

Plate 4.41 F4 folds deforming S3, Westport Grit Formation, Westport (M 0205 8453).

contained within the pluton. This fabric therefore progressively changes orientation from the northwest side of the pluton, where it has a mean orientation of 054/57, to the southeast side of the pluton where it trends northeast- southwest and has a sub-vertical attitude.

This earlier fabric is succeeded by steeply inclined discrete mylonitic zones that are concentrated at the contacts between lithological units within the pluton. These consist of fine-grained zones which cut the main fabric and shear-bands. Both quartz and feldspar show grain size reduction in these narrow zones which although narrow and relatively uncommon, represent significant displacement, (Mc Caffrey, pers comm).

In the Attiappleton area, at the southwest of the pluton, the initial sinistral fabric is extensively overprinted by moderately inclined, penetrative dextral ductile shear fabrics with a mean orientation 193/58W that result from movement on the Knockaskibole fault. Later brittle deformation produced a series of shallowly dipping discrete faults, thrust slices and duplexes (Plate 4.42).

4.9.2 Lough Talt Adamellite

The Lough Talt Adamellite contains two tectonic fabrics that are highly discordant with the S3 fabric that is developed in the country rocks. The earliest of these is a sinistral late-to post-crystallization fabric (Hutton, 1988) that has a mean orientation of 045/50. It consistently dips moderately to the southeast, in contrast to the steep northwest dips of the S3 foliation in the metasediments immediately southeast of the contact. This indicates that the fabric was produced by a different deformation. The origin of this fabric is discussed in section 5.5. The pluton fabric is succeeded by discrete dextral shears and fractures generally with a spacing of <10cm, that reflect minor late dextral deformation.

Plate 4.42 Contractional imbricate stack developed in the Ox Mountains Granodiorite at the northwest end of the Knockaskibole fault zone (G 1900 0565).

4.9.3 The Lough Easky Adamellite

The Lough Easky Adamellite is deformed by two tectonic fabrics that are correlated with those that deform the LTA. The first fabric is a preferred dimensional orientation of quartz and feldspar. The quartz shows ductile deformation and augens feldspar megacrysts that have been fractured by brittle deformation mechanisms. This fabric, which is less strongly developed than in the LTA, is best observed in the margins of the pluton (G 4480 2280) and consistently indicates sinistral shear. Offset s of brittle deformation consistently indicate a sinistral component of shear. Andrews (1984) determined the shear sense by examining the offsets of fractures within individual feldspar grains. These fractures have developed to accommodate the ductile deformation within the matrix, and thus have dextral offsets which are antithetic to the overall sinistral shear sense. The orientation of the fabric in the pluton (030/64) is similar to that of the S3 foliation in the country rocks. In common with the LTA, this penetrative fabric is succeeded by discrete dextral shears. (Mc Caffrey, pers comm, 1988).

4.10 CORRELATION OF THE DEFORMATION CHRONOLOGIES OF THE OX MOUNTAINS SUCCESSION AND CLOONYGOWAN FORMATION.

This correlation is of critical importance to an understanding of the tectono–thermal history of the inlier. The contacts between the two units are not exposed in either areas 6.a, 6b or 6.c. The structures must therefore be correlated on the basis of their geometry and relationship to the regional metamorphism. The absence of unequivocal cross-cutting relationships between the fabrics in these units has led to a number of differing interpretations. Taylor (1969), and Long and Max (1977) regard the Cloonygowan Formation as having been deformed synchronously with the Ox Mountains Succession. Phillips et al (1975) and Andrews et al. (1978) state that the first fabric which they recognise in the Cloonygowan Formation (Sc1) (Sc2 in the present chronology) can be correlated with D4 cataclasis in the rocks of the Ox Mountains Succession. They therefore regard the deformation of the Upper Dalradian rocks to be the earliest Caledonian event and to overprint the earlier deformation within the Ox Mountains Succession which they consider to be pre–Caledonian.

The present work (summarized in Figure 4.21) demonstrates that the deformation history in the Upper Dalradian rocks of the Cloonygowan Formation is considerably more complex than was previously recognised, and that both areas share a common D1-D4 polyphase deformation history. The correlation of Phillips et al (1975) between Sc1 (Sc2 present chronology), which is represented by a shallowly dipping penetrative fabric inclined at a small angle to bedding, and D4 (Sc3 in the present chronology), a sub-vertical S-C fabric, is not accepted by the present author. The Sc1 fabric of Phillips et al. (1975), (Sc2, present chronology) does however have an identical geometry to the S2 fabric in the higher grade rocks of the Ox Mountains Succession. This deformation is succeeded by a phase of deformation that produces a steeply inclined fabric in both sets of rocks (D3, present chronology), and a phase that produces conjugate folds with gently plunging axes and shallowly inclined axial planes (D4). Therefore there is no geometrical basis for regarding the deformation in the Cloonygowan Formation as later than and overprinting the deformation in the higher grade rocks of the Ox Mountains Succession.

The geometrical similarity between the deformations of both areas strongly suggests that they have been deformed synchronously by the same orgenic episode. If however this correlation is accepted the differences in metamorphic grade experienced by the two areas must be accounted for.

West of the Knockaskibole fault the pelites of the Ox Mountains succession lie in the garnet zone of the greenschist facies and contain the assemblage (gar-

	CLOONYGOWAN FORMATION	OX MOUNTAINS SUCCESSION
D1	S1 weak bedding-parallel alignment of musc-chl	S1 penetrative fabric observed in hinges of F2 folds
D2	S2 penetrative fabric inclined at a small angle to bedding F2 close to isoclinal folds D2 displacement on Cloonygowan Fm - Ox Mts. Succn. contact indicated by D2 shear bands observed in area 6b.	S2 strong penetrative fabric now largely transposed into composite S1-S3 fabric F2 rare F2 folds preserved in hinges of F3 structures
D3	S3 discretely spaced pressure solution fabric subvertical, intersects S2 at a high angle F3 upright, isoclinal folds, axes oriented parallel to X2 SC3 Shear bands developed in area 6a adjacent to contact with Ox Mts Succn F4	S3 generally steeply inclined spaced fabric, intersects S2 at a high angle tectonic slides developed F3 upright, tight to isoclinal folds parallel to gently plunging stretching lineation SC3 Shear bands extensively developed throughout the Ox Mts. Succn.
D4	Open folds, axes parallel to X3. S4 Gently inclined axial planes, no axial planar fabric.	F4 open folds, axes parallel to X3 S4 Axial panes gently inclined, no axial planar fabric
D5		Renewed sinistral transcurrent deformation affects adamellites, succeeded by minor discrete dextral shears.

FIG 4.21

net, biotite, chlorite, chloritoid, chlorite + muscovite, quartz and plagioclase). Garnet is however absent from the pelites of the Cloonygowan Formation of area 6a, and the assemblage here (quartz, albite, muscovite, chlorite) indicates that the metamorphism there has not exceeded the chlorite zone of the greenschist facies. There is therefore a significant difference in metamorphic grade between these two areas. The contrast between the metamorphic grade of the Cloongowan Formation of area 6b and the adjacent Ox Mountains Succession appears to be greater. The pelitic rocks of the Ummoon Formation of the Ox Mountains Succession 1000m northwest of the Cloonygowan Formation contain garnet, staurolite, biotite, muscovite, plagioclase + quartz, indicative of metamorphism under staurolite zone conditions of the amphibolite facies, whereas the rocks of the Cloonygowan Formation have not exceeded the chlorite zone of the greenschist facies.

This significant contrast in metamorphic grade must be accounted for by juxtaposition following MP3 when the metamorphic grade still lay at the amphibolite grade. In the case of the Cloonygowan Formation of area 6a this is likely to have been accomplished by net vertical displacement on the Callow Fault located in the unexposed ground that separates the two units. At Ardvarney (area 6b) in the southwest of the inlier, the smaller relative vertical displacement required must also have occurred by late or post D3 movement on the unexposed contact that separates the two units.

CHAPTER 5

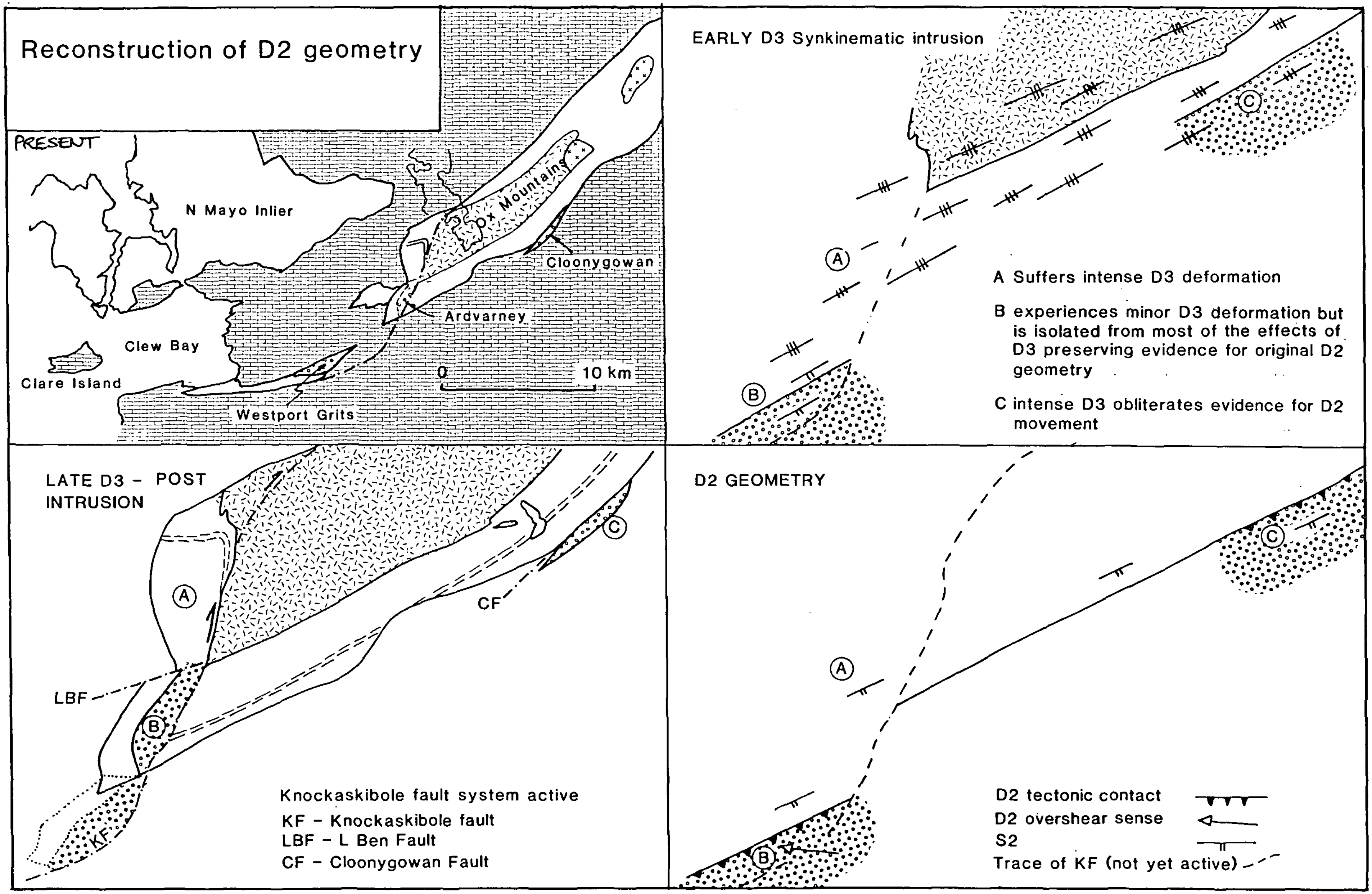
TECTONOTHERMAL DEVELOPMENT OF THE INLIER.

The aim of this chapter is to integrate the structural, kinematic and metamorphic data of chapter 4 into a model for the development of the inlier. No attempt is made to illustrate stratigraphic boundaries on figures 5.2–5.8, as this would considerably complicate the diagrams and conceal important structural relationships. The relationship of the tectono-metamorphic model to regional tectonics and the implications that it has for terrane accretion models of the Highland Boundary Fault in W. Ireland is examined in chapter 6. The information that the model provides on the general problem of processes and deformation geometries in mid-crustal shear zones is examined in chapter 7.

5.1 D1 : THE FIRST DEFORMATION

The original geometry of S1 is unknown, so it is therefore not possible to determine the geometry and kinematic significance of D1. The fine grained nature and composition of the S1 fabric, which is expressed as a rarely preserved inclusion trail of MS1 muscovite and quartz in albite porphyroblasts suggests that the metamorphic grade lay in the low greenschist facies prior to D2, (Figure 5.9).

The absence of a clearly defined S1 fabric from areas within the Ox Mountains that have experienced a relatively weak D2 deformation, suggests that S1 was originally developed at a small angle to lithological banding, as in these areas S2, which itself is consistently at a small angle to, or sub-parallel with bedding, is considered to be insufficiently well-developed to have completely transposed a pre-existing fabric that was originally inclined at a high angle to lithological banding. No evidence has been observed to suggest that any major structures



Reconstruction of D2 geometry

EARLY D3 Synkinematic intrusion

A Suffers intense D3 deformation

B experiences minor D3 deformation but is isolated from most of the effects of D3 preserving evidence for original D2 geometry

C intense D3 obliterates evidence for D2 movement

LATE D3 - POST INTRUSION

Knockaskibole fault system active

KF - Knockaskibole fault

LBF - L Ben Fault

CF - Cloonygowan Fault

D2 GEOMETRY

D2 tectonic contact

D2 overshear sense

S2

Trace of KF (not yet active)

FIG 5.1

related to D1, such as the major D1 tectonic slide developed in the Dalradian of south Donegal (Central Donegal Slide, Alsop 1987), are present in the area.

5.2 D2 : THE SECOND DEFORMATION

Over most of the inlier D3 deformation has transposed S2, thereby obscuring the original geometry and kinematics of D2. However D2 was clearly an important phase in the tectonic history of the inlier, since it produced a strong penetrative fabric and F2 folds. Evidence preserved at Ardvarney demonstrates that the contact between the Cloonygowan Formation and the Ox Mountains Succession was active at this stage. In this section the geometry and kinematics of D2 structures are reviewed, and by removing the effects of subsequent deformation an attempt is made to reconstruct the D2 geometry of the inlier and the original geometry of the contact between the Cloonygowan Formation and the Ox Mountains (Figure 5.1).

5.2.1 Fabric development and geometry (Figure 5.2)

D2 deformation has resulted in a penetrative fabric in all the units of the Ox Mountains. Where lithological banding is observed this fabric always lies parallel to it. Unfolding the D3 deformation suggests that S2 was originally shallowly inclined. The 'type three' refolding pattern (Ramsay 1967) formed where F2 and F3 folds interfere suggests that F2 fold axes originally had a northeast-southwest orientation. This, combined with the largely linear outcrop pattern suggests that the strike of S2 was likely to have been northeast-southwest.

While small scale F2 folds are occasionally observed, the absence of more widespread vergence information including cleavage–bedding relationships means that the existence of larger scale F2 folds can only be inferred from stratigraphic data, where the observed stratigraphic configuration cannot simply be accounted

D2 - PENETRATIVE FABRIC DEVELOPMENT

Cloonygowan Formation thrust into contact with the Ox Mountains Succession

Lough Anaffrin antiform develops

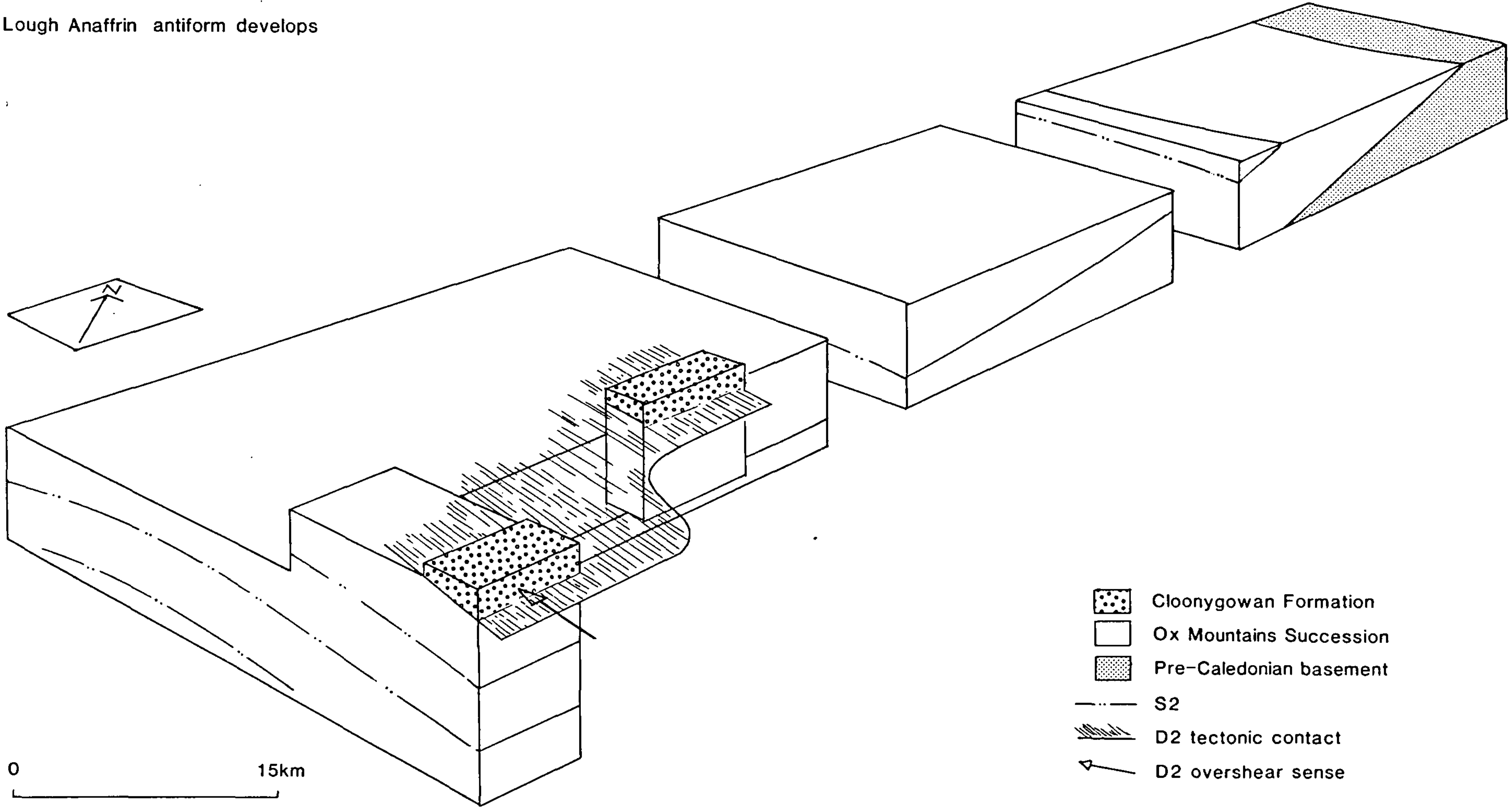


FIG 5.2

for by D3 or subsequent deformations. The only major D2 structure identified in this way is the Lough Anaffrin Antiform, (section 4.3.3). The observed stratigraphy and the rarity of mesoscopic and microscopic F2 folds is consistent with the absence of further major F2 folds.

5.2.2 Tectonic contacts

The intensification of S2 and the progressive development of shear bands towards the contact between the Cloonygowan formation and the Ox Mountains Succession in the Ardvarney area strongly suggests that this is a tectonic contact of D2 age. Following the juxtaposition of the Cloonygowan Formation rocks against those of the Ox Mountains Succession (Figure 5.1), the inlier was affected by D3 transpressive deformation that refolded lithological banding and S2 and produced a strong generally steeply inclined fabric with a northeast–southwest strike, at a high angle to S2 in all areas apart from Tullycommons and Ardvarney (see sections 4.7.3 and 4.8.2).

It is evident that if the displacement on the Knockaskibole fault is restored, the areas in question lay approximately 10 km to the south of their present location. This horizontal displacement alone may have removed the areas sufficiently far from the locus of the main D3 deformation to account for the absence of D3 deformation within them. In addition the area may also have been vertically separated from the remainder of the Ox Mountains Succession by one of two mechanisms:

1. A low angle detachment, that presently lies below the current exposure level may separate the two areas. This would explain the very abrupt difference in the deformation of the two areas.

2. There may be considerable vertical displacement on the east-west trending late fault (The Lough Ben Fault) that presently separates the two areas. Stratigraphic data discussed in section 2.1.4 suggest that the displacment on this fault may be

large and that the southern block may have been downthrown from a significantly higher level than the northern one.

Whilst the first of these models would provide a very efficient mechanism for the partitioning of deformation that is observed, it cannot however be verified as such a detachment would lie below the present exposure level. While the second mechanism undoubtably occurs, it is unclear whether it alone could account for the difference in structural histories of the two areas.

5.2.3 Comparison with the contacts at area 1a : Evidence for a D2 contact

Structural data indicate that the contact between the Ox Mountains Successionb and the Cloonygowan formation was active as a gently plunging, steeply inclined D3 sinistral shear zone. Structural relationships observed at Ardvarney (Sub–area 6b) suggest that the original contact between these two units may have also had an important D2 tectonic history. The contact between the two units is now expressed as a steeply inclined northeast–southwest trending fault (The Callow fault). The Cloonygowan Formation and the Ox Mountains Succession must have been juxtaposed following MP3 when the Ox Mountains Succession rocks were still undergoing amphibolite facies metamorphism, as metamorphic grade in the Cloonygowan Formation never exceeded the greenschist facies. This juxtaposition is likely to have been achieved by a combination of late D3 sinistral movement on the gently northeast plunging shear zone, and most importantly by net vertical displacement on the Callow fault.

5.2.4 Evidence for D2 displacement on other tectonic contacts

S2 generally lies parallel to lithological banding and therefore to the boundaries between lithological units. These boundaries, principally the Leckee

Quartzitic Formation - Ummoon Formation and, Ummoon Formation - Lower Lismoran Formation, are also rheological boundaries and are the sites of D3 tectonic slides. The S2 foliation would therefore appear to have a favourable orientation to act as a plane of movement. Unfortunately, due to the intense D3 deformation, it is not possible to determine if they have also localized strain during D2. It is likely that they acted as tectonic boundaries during D2 although no direct evidence is available for this.

5.2.5 Reconstruction of D2 geometry (Figure 5.1)

The present overshear sense observed at Ardvarney (240°Azimuth) is unlikely to represent the original geometry of the contact. The strata of the Tullycommons area have clearly been reorientated, so that they now lie perpendicular to regional strike. Figure 5.1 illustrates the simplest reconstruction, which involves restoring the movement on the Knockaskibole Fault. This restoration must assume the geometry of the Knockaskibole fault southeast of Newantrim where it is concealed beneath Carboniferous cover. The interpretation used in Figure 5.1, that the fault has a sigmoidal trajectory and anastamoses into the major fault system on the south side of Clew bay, is consistent with the observed magnetic data (Max et al., 1983) which show that the area is dominated by linear northeast–southwest trending magnetic features. Restoration of the movement on the Knockaskibole fault produces a 35° clockwise rotation of strike of the rocks of the Tullycommons area, and a corresponding reorientation in the overshear direction of the overlying Cloonygowan formation. The original geometry of the contact inferred by this reconstruction is that of a thrust directed towards 275°azimuth.

5.3 D3 : PROGRESSIVE DEVELOPMENT OF A MID-CRUSTAL STRIKE-SLIP FAULT SYSTEM

The present geometry of the Ox Mountains results largely from progressive D3 sinistral transpressive deformation. This produced a kilometric scale mid-crustal strike-slip fault system, whose development is described below. The terminology used in figures 5.3–5.7 follows the work of Woodcock and Fischer (1986) on strike slip systems. This terminology is used and developed further by the present author.

5.3.1 D3 A : Pre-intrusion geometry - shear zone initiation (Figure 5.3)

Evidence that the granodiorite was syn-kinematic with respect to D3 is outlined in chapter 4. Tight-isoclinal F3 folds are cut by granodiorite sheets in the metasedimentary envelope and F3 folds with a more open geometry are preserved in rafts within the granodiorite where they are also commonly cut by granodiorite sheets and veins. Often in the latter situation the granite veins are themselves folded but less tightly than the metasediments which they intrude. Data preserved in rafts within the pluton suggest that it was intruded into the core of a northeast-southwest trending F3 antiform, and the orientiation of F3 folds, parallel to the northeast-southwest trending regional stretching lineation suggests that this structure developed in response to transcurrent deformation. The metamorphic grade at this time still lay at the amphibolite facies following the MP2 peak of regional metamorphism.

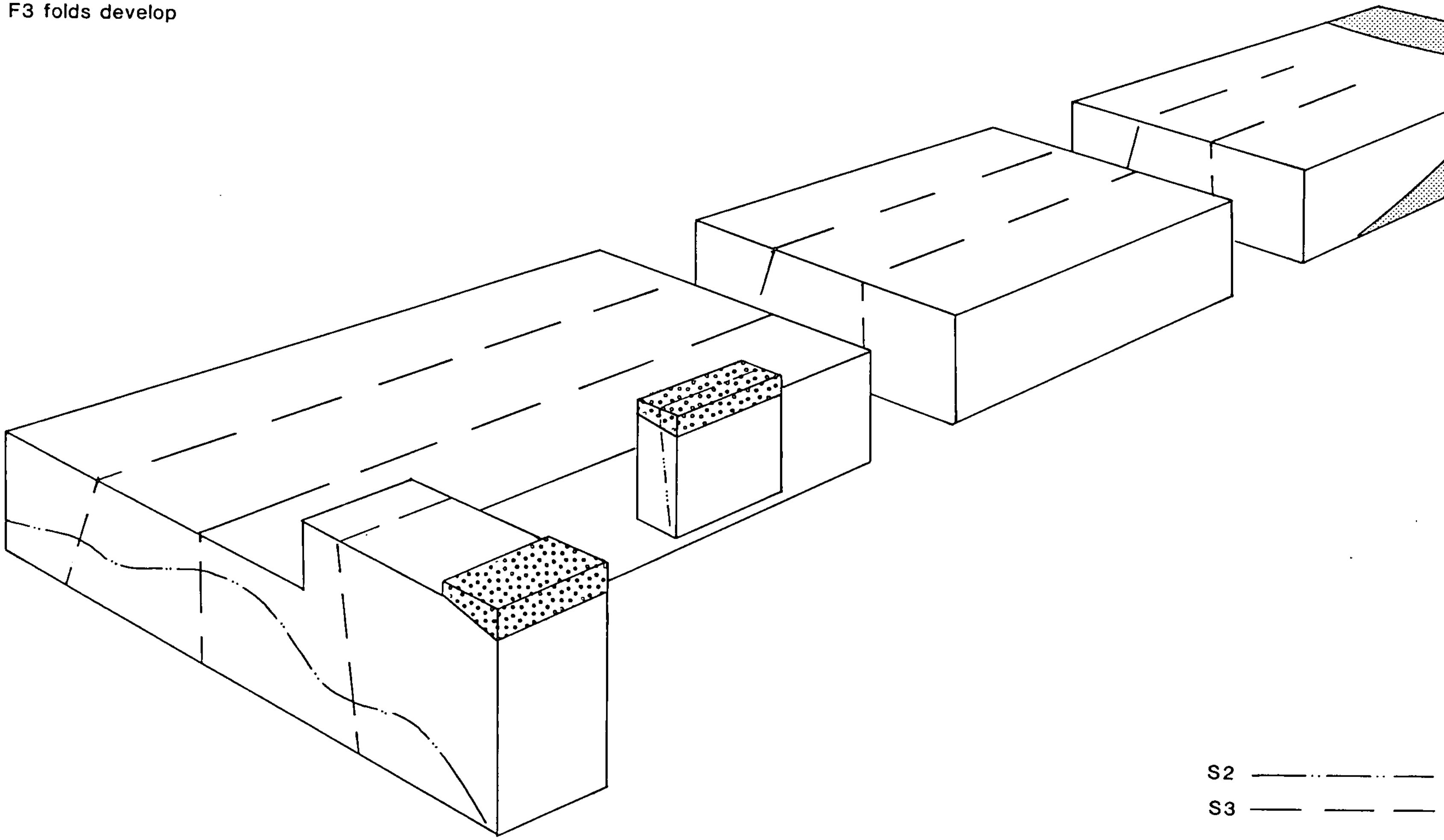
At this stage the Cloonygowan Formation of areas 6a and 6c was undergoing deformation with a similar geometry but at a structurally higher level. The Cloonygowan formation rocks of area 6b (Ardvarney) were at this stage isolated from the D3 deformation that affected the remainder of the inlier.

FIG 5.3

D3 A PRE INTRUSION GEOMETRY

S2 folded

F3 folds develop



5.3.2 D3 B : Synkinematic intrusion - shear-zone active (Figure 5.4)

Syn-kinematic intrusion of the granodiorite followed the initial folding and fabric development, described in D3 A. Geobarometric data (Yardley and Long, 1981), indicate that intrusion occurred at a pressure of 6–7kbar which corresponds to an intrusion depth of approximately 24km. This is consistent with the absence of a well developed thermal aureole, which suggests that intrusion occurred close to the amphibolite facies peak of regional metamorphism, and that the metasedimentary envelope was hot and unresponsive to further heating.

Small-scale examples of intrusive relationships observed on the northwest side of the pluton at Pontoon suggest that the emplacement of sheets occurred in a sinistral dilational environment (Mc Caffrey, pers comm). These data may conflict with the widespread evidence in the country rocks that the ambient tectonic environment immediately before and synchronous with the emplacement of the pluton was predominantly transpressive. A number of possibilities exist that may account for this:

a) Emplacement could have been achieved in a transpressive regime only if the magma buoyancy force exceeded the ambient transpressive strain, i.e. emplacement would be forceful. Emplacement by such a mechansim would result in increased ballooning strains in the country rocks, concentrated at the margin of the main body of the pluton and the margins of the smaller sheets of granodiorite in the contact zone. No evidence has been seen to support this. While it is possible that the intense post–intrusion transpressive deformation may have modified such ballooning strains so that they are no longer apparent, this is considered to be unlikely as the later deformation is not considered to be sufficiently intense to have obliterated evidence for ballooning strains everywhere had they been present.

b) It is probable that the shear-zone, while dominantly transpressive in nature is likely to have developed by the operation of a series of non–parallel high strain

D3 B – SYNKINEMATIC INTRUSION

Sinistral shear zone

Intrusion of OX Mts Granodiorite into core of tightening F3 antiform

S3 fabric develops in Ox Mts Succn and Coonygowan Fm

S2 transposed

F3 folds parallel to NE–SW stretching lineation

Spacing of lines reflects intensity of S3

Ardvarney area 6b isolated from the effects of D3

Maximum PT conditions within the Ox Mountains

Succession – Amphibolite facies

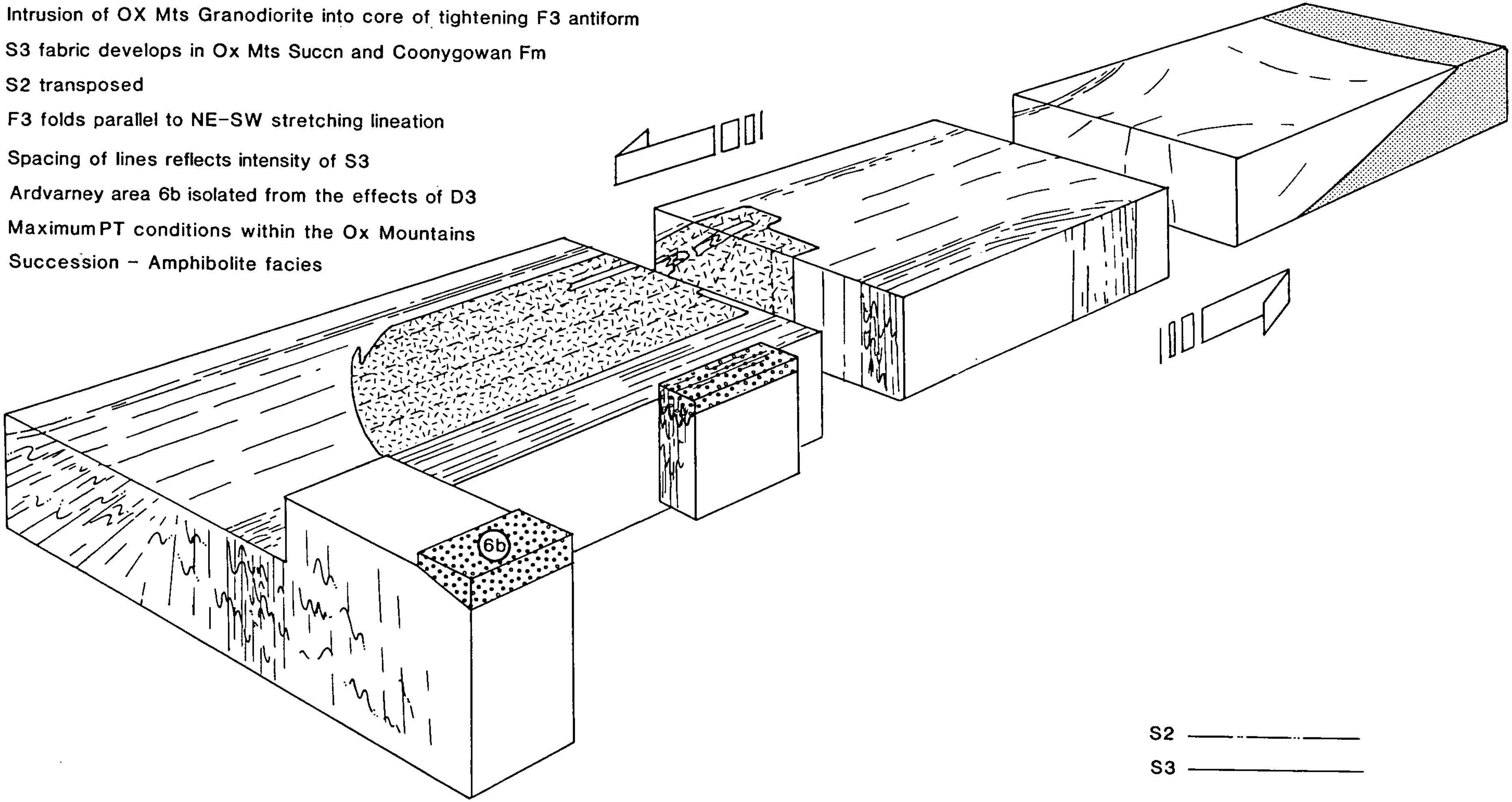


FIG 5.4

zones, active at different rates at different times. This inhomogeneous deformation will result in the development of strain gradients and hence displacement gradients within the metasedimentary envelope. This will be reflected by areas of high transpressive strain separated by areas of low transpressive strain. The pluton may thus have exploited a zone of diminished transpressional strain or even dilational strain within an overall transpressive shear zone. If this were the case then the pluton now precisely occupies the site of the dilation, as there is evidence for dilational strains in the country rocks.

Emplacement could therefore be achieved by intruding a coalescing series of sheets that enclose F3 folds, and create roof septa, pendants and raft trains. This model is supported by the mapping of McCaffrey who noted that the separate bodies of differing igneous lithotype, intruded on the northwest side of the pluton at Attymass (G 3125 1250) were commonly separated by rafts and sheets of metasediment. Following the initial intrusion the presence of magma within the shear zone would have further increased the anisotropy and therefore strongly influenced the subsequent development of the structure.

5.3.3 D3 C : Braided fault system develops in response to continued sinistral shear (Figure 5.5)

Transpressive deformation that originated earlier in D3, continued synchronously with intrusion and subsequent cooling of the pluton. This deformation produced a steeply inclined crystal-plastic fabric within the pluton and the metasedimentary envelope. The deformation locally intensifies to form a series of high-strain zones which converge both laterally and vertically to give rise to a braided strike–slip fault system. Development of this structure began when metamorphic conditions were at the amphibolite facies and continued during the cooling and retrogression that preceeded D4, (Figure 5.9). The essential components of this structure are identified and described below, in terms of such a

D3 C – SINISTRAL SHEAR ZONE ACTIVE

S3 intensifies

Discrete zones of movement develop

shear bands develop throughout the inlier

PT conditions – amphibolite facies

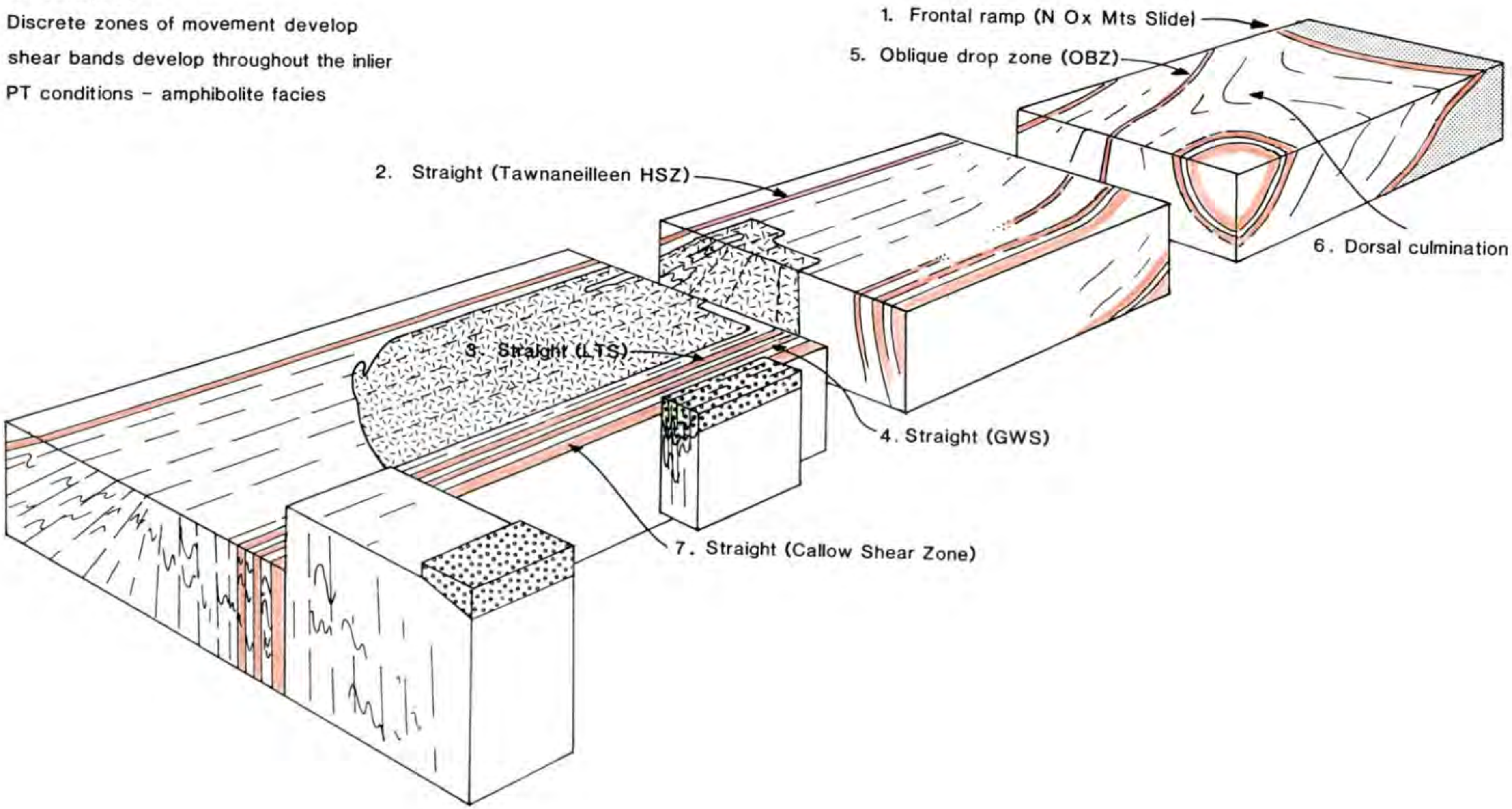


FIG 5.5

braided fault system. The likely kinematic history of the entire structure is then discussed. Numbers in the text in brackets refer to figures 5.5-5.8.

(1) N. Ox Mountains Slide (Frontal ramp). This basement-cover interface is the deepest structural level observed in the inlier. The pre-Caledonian granulite facies psammites of the northeast Ox Mountains, which dip gently to the south behaved as a rigid block that controlled the structure of the metasediments of the Ox Mountains Succession as they are driven over it from the southwest. This produced a strong northeast–southwest trending stretching lineation that plunges gently to the southwest, and a bend segment in the Callow Shear Zone (7). The structure therefore has the geometry of a frontal ramp in a thrust system, as it is inclined gently southwest and directly confronts the overthrust vector. It should be emphasised that this is likely to be simply a reflection of the basement topography and is distinct in origin from the 'propagated ramps' of thrust systems.

(2) Tawnaeilleen high strain zone (THSZ)(Straight). Located along the northwest side of the inlier the TSZ (2), has the geometry of a straight or fault segment that lies parallel or sub-parallel to the regional slip vector. This zone of high strain takes up considerable strike slip motion on the northwest side of the inlier.

(3) Lough Talt Slide (LTS) (Straight). Unlike the TSZ this slide is oblique to the stratigraphy. It cuts out the Leckee Transition Member southwest of Lough Talt (see section 4.3.5.1, Figure 4.10) thus bringing the Ummoon Formation into contact with the Leckee Quartzite in the rest of the inlier to the southwest.

(4) Glennawoo Slide (GWS) (Straight). This lies parallel to the stratigraphic boundary between the Ummoon and the Lower Lismoran Formations, and consists of a 'straight' section (4a) and a 'bend' section (4b). The change in orientation of the structure from straight to bend, is likely to be in direct response to the buttressing effect of the frontal ramp (1).

(5) Oblique Drop Zone (ODZ). This separates the steep zone to the west of Lough Easky from the shallow zone to the east, and accommodates part of the relative vertical displacement between two blocks, the eastern of which probably underwent more rapid uplift as a result of being driven up over the frontal ramp (1). This greater uplift to the east may be reflected in increased intensity and scale of D4 vertical shortening structures. The ODZ may represent a splay from the GWS. The ODZ also appears to be the locus for the emplacement of the later Lough Easky Adamellite.

(6) Dorsal Culmination. This could have developed in response to driving the metasediments of the Ox Mountains Succession east of the ODZ (5) over the frontal ramp (1), either due to; 1. butressing, as resistance to further thrusting over the basement builds up or 2. the geometry of the basement topography being curvi linear.

(7) Callow Shear Zone (CSZ). This separates the Upper Lismoran Formation from the Cloonygowan Formation and is the broadest zone of high strain in the inlier. With the geometry of a straight segment, movement on this structure, recorded by the development of sinistral shear bands in both the Cloonygowan Formation and the Upper Lismoran Formation, as well as subsequent faulting, has brought the Cloonygowan Formation into contact with the Ox Mountains Succession. This contact is therefore one of the most important in the inlier.

5.3.4 D3 D : Lock–up and break-back development (Figure 5.6)

Following the sinistral transcurrent deformation that produced movement on structures (1-7), the major fault strands now locked up and became inactive. The westerly dipping Knockaskibole fault (8), comprising a break-back (8a) and a contractional imbricate fan (8b) may then have developed as a response to continued transcurrent deformation. The fault displaces, and therefore clearly post-dates movement on the high strain zones (GWS and LTS) developed on the southeast

D3 D -

Knockaskibole break back and contractional imbricate active as dextral displacements

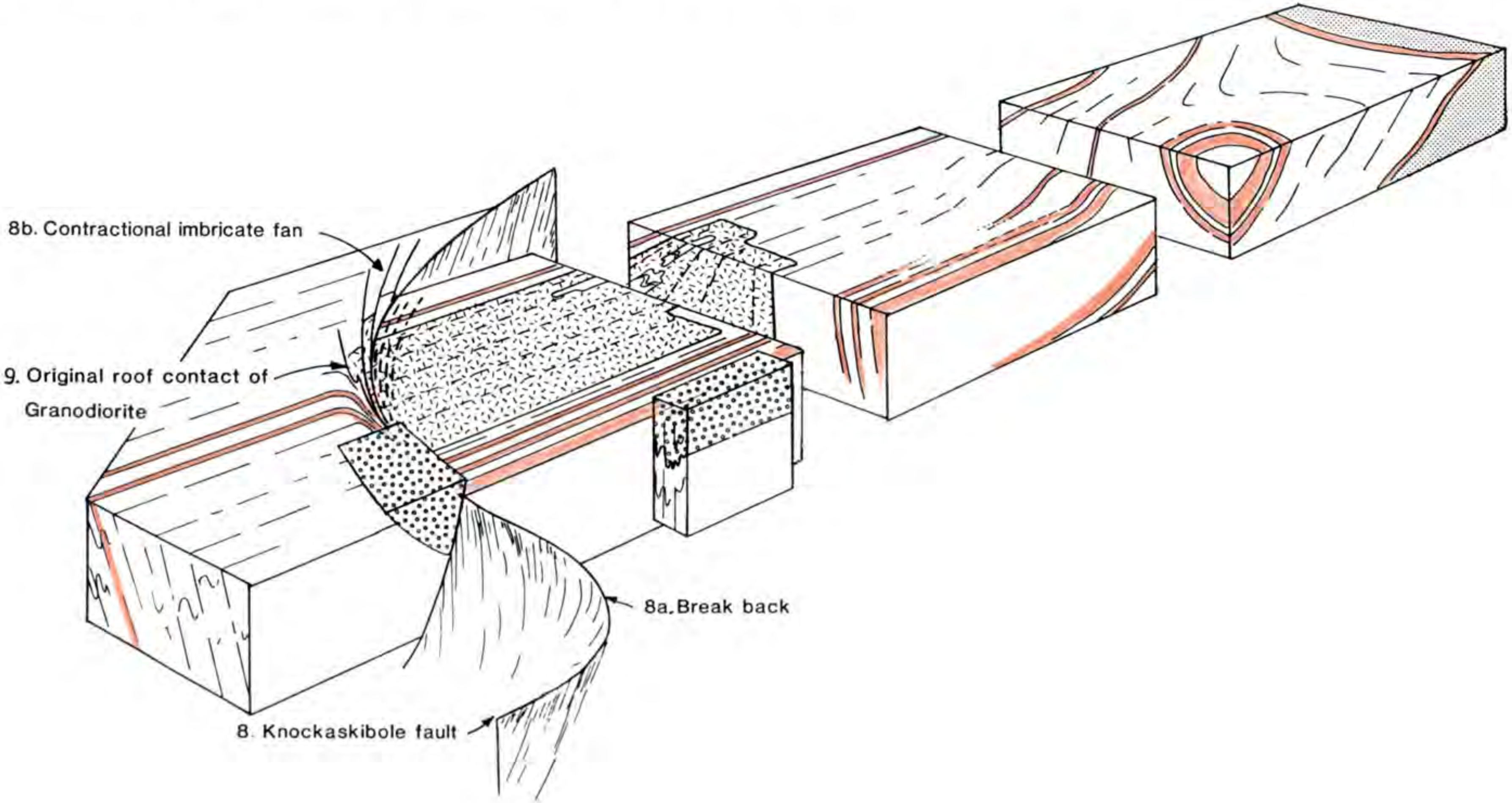


FIG 5.6

side of the inlier. Its timing, and basic southwest inclined geometry and movement sense are analogous to that of a break-back thrust (Butler, 1987), (a structure that occurs at a structurally higher level and behind the previously developed structure, in this case the frontal ramp (1)). At its northern end the fault divides into a number of closely spaced fault splays, each with dextral displacement, the most westerly of which isolates the original intrusive roof contact of the granodiorite (9). The other wedges are of granodiorite which have strong late dextral fabrics imposed upon earlier sinistral fabrics associated with the main sinistral shearing that affected the inlier. The geometry of these wedges suggests that they are contractional imbricates (Butler, 1987) that developed late in the sequence of deformation, when the shear-zone locked up and further movement on the basement cover interface was energetically unfavourable.

5.3.4.1 Kinematic relationships - development of a linked strike-slip fault system: Figure 5.5 illustrates that the geometry of the inlier is controlled by a series of strike slip faults that anastamose both laterally and vertically to form a complex braided fault system. At least in the northeast end of the inlier this is a response to basement topography. In this section the kinematic relationships of the various components of the structure are examined, and by viewing the structure as a simple package of Dalradian stratigraphy that is driven over a gently south dipping basement–cover interface by sinistral tranpressive deformation, the likely sequence of movement on each of the fault strands can be discussed.

Initial folding and fabric development throughout the inlier was followed by intrusion of the granodiorite and the development of more discrete zones of deformation. These structures numbered (1-7) (figure 5.5) take up the greater part of the sinistral displacement that occurs during D3. Sinistral transpression dictates that the straight segments (4 and 7) were activated early, as they are deformed during the ramp climb over the frontal ramp (1). As deformation continued, the dorsal culmination (6) developed to accommodate material driven up over the south-dipping basement–cover interface and continued sinistral displacement was taken up on the remaining northeast southwest orientated straight segments (2 and 3). The oblique drop zone decoupled the area south of the frontal ramp, which was undergoing rapid uplift and retrogression, from the area to the west, which is dominated by moderately-steeply inclined linear fabrics associated with the transcurrent deformation (2). Deformation then continued under conditions of decreasing metamorphic grade. Sinistral displacement occurred on the straights (2,3,4 and 7). Displacement may have occurred simultaneously on all of these structures or may have shuffled between them with different structures taking up the displacement at different times. The amount of displacement represented by these structures is considerable. However, as the shear-zone lies parallel to the regional strike it is not possible to use bedding offsets to determine the displacement. Since the southwest extent of the straight segments is concealed beneath post-metamorphic cover, the total length of the zone is not known, although it clearly exceeds its exposed strike length of 60 km.

Eventually, late in D3, sinistral displacement ceased on the straight segments, probably due to lock up or buttressing that occurred following increased resistance to further displacement on the frontal ramp. At this point the Knockaskibole fault developed as a break–back cutting up section and above and behind the frontal ramp to breach (Butler, 1987) the pre-existing straights. Movement on this structure has produced a pervasive ductile dextral fabric in the granodiorite, that overprints the earlier sinistral fabric, indicating that deformation occurred under greenschist facies conditions in contrast to the amphibolite facies deformation that produced the earlier sinistral fabric in the granodiorite (Mc Caffrey, pers comm). Wedge shaped segments of the granodiorite are isolated by splays of the structure giving the geometry of a contractional imbricate fan. Displacement on this structure is closely constrained by the offset of the straight segments of the shear-zone (3,4 and 7) which indicate a dextral component of about 10km.

5.4 D4 : REGIONAL UPLIFT (Figure 5.7)

D4 structures reflect the fundamental change in kinematics that followed D3 and the intrusion of the Ox Mountains Granodiorite, when transcurrent deformation ceased and uplift began throughout the inlier. D4 is everywhere represented by folds that have an angular or chevron geometry. The axes of F4 folds lie parallel to or at a small angle to the well-developed L3 stretching lineation. S4 axial planes are consistently shallowly inclined (generally <30°) and often form conjugate sets. F4 folds have therefore clearly formed in response to vertical shortening with an associated strike–perpendicular extension. The absence of a penetrative axial planar fabric associated with F4 folds which simply bend MS3–MP3 micas and produce undulose extinction and some recrystallization of quartz confirms that metamorphic conditions lay in the low greenschist facies during D4 (Figure 5-9). D4 structures are developed throughout the inlier in both the Ox Mountains Succession and the Cloonygowan Formation. They find their greatest development southeast of the Lough Easky Adamellite. In this area, which corresponds to an area of strong retrograde metamorphism, D4 structures have an amplitude of up to 100m and commonly exceed 10m, which is an order of magnitude greater than that observed elsewhere in the inlier. In this area at least, they represent a significant amount of shortening, probably up to 40%.

D4 structures could at least in part reflect the vertical shortening that may have accompanied the uplift that occurred between the intrusion of the Ox Mountains Granodiorite and the intrusion of the Lough Talt and Lough Easky Adamellites. The coincidence of the timing and development of these structures with the PT history of the inlier suggests a causal relationship between them. The alternative possible explanation that they could have been produced due to footwall loading under a now totally eroded thrust sheet, is considered to be unlikely as no evidence for such a large thrust sheet has been found. It is also interesting to

D4 REGIONAL UPLIFT

Accommodation structures developed

F4 fold axes orientated parallel to X3

PT conditions - Low greenschist facies

Cloonygowan Fm finally juxtaposed with Ox Mountains Succession

along LBF & CF syn or post D4

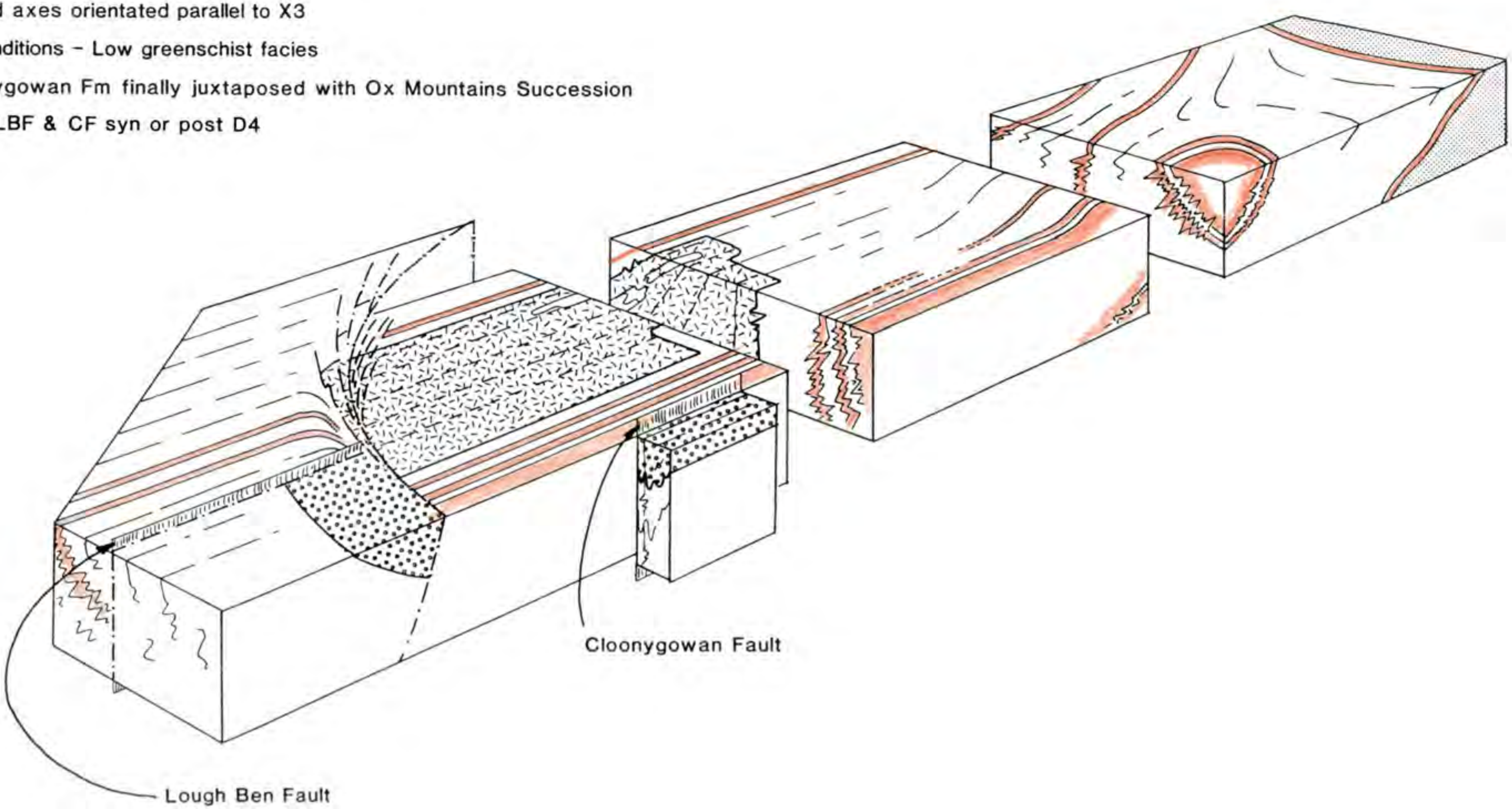


FIG 5.7

note that identical structures are seen in the Main Donegal shear zone (Hutton, 1977; author's own observations).

5.5 D5 : RENEWED TRANSCURRENT DEFORMATION; EMPLACEMENT AND DEFORMATION OF THE ADAMELLITES (Figure 5.8)

The LEA is located on a structure that is termed an oblique drop zone (Figure 5.8) that was active during D3, and separates two large blocks that are likely to have experienced different magnitudes of vertical movement. The elliptical shape of the adamellite is well constrained, and its elongation parallel to the ODZ is consistent with the earlier D3 structure having exerted a fundamental control on its emplacement and subsequent deformation. The penetrative fabric occurring throughout the pluton has a similar orientation to the D3 fabric that developed earlier in the metasedimentary envelope and consistently indicates sinistral shear.

The Lough Talt Adamellite (LTA) is located immediately north of the northeast limit of the Ox Mountains Granodiorite, 750 m northwest of the Glennawoo Slide. The shape and extent of the pluton is very poorly constrained as only its northwest and southeast contacts are exposed. The presence of a shallowly southwest dipping floor contact between the adamellite and the country rocks suggests that the pluton has been intruded, at least here, upwards from the southwest. The absence of data on the shape of the pluton and the geometry of its contact with the Ox Mountains Succession to the southwest precludes any further conclusions to be drawn. A strong sinistral S/C fabric is developed in the pluton. This fabric shares a common strike with the fabric in the country rocks, but dips to the northwest rather than to the southeast, and cross-cuts the contact between the pluton and the country rocks. The development of a sinistral S/C fabric oblique to the contact with the metasediments in both plutons, suggests that the fabrics

D5 – ADAMELLITE INTRUSION

Renewed sinistral shear

Sinistral shear fabrics in LEA & LTA

PT conditions – Low greenschist facies

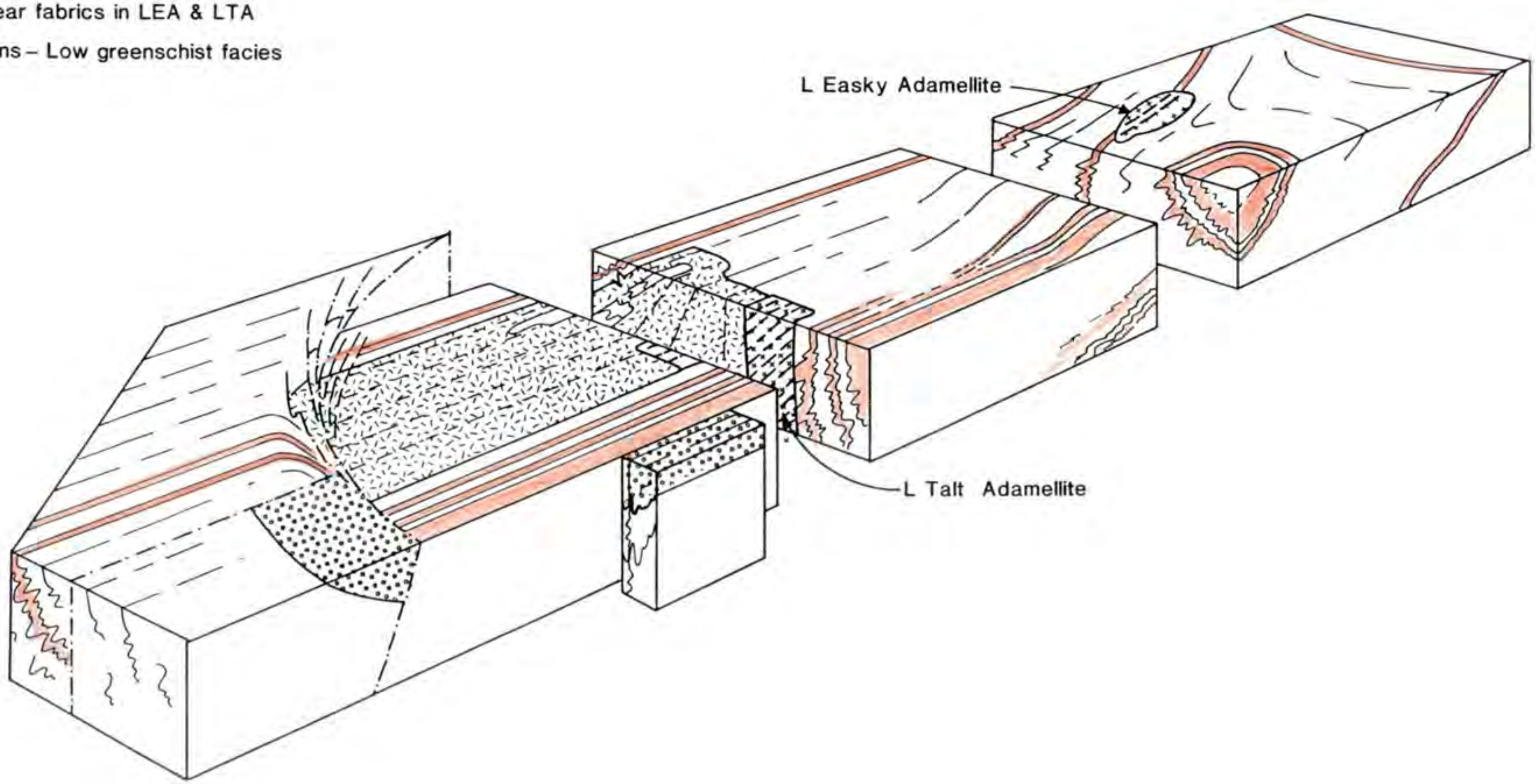


FIG 5.8

are of tectonic origin rather than simply due to the emplacement of the pluton. These data strongly suggest that the inlier was subjected to a final period of sinistral transcurrent shear, shortly following the emplacent of the two plutonic bodies at approximately 401±33 Ma. The relatively weak nature of the deformation in the adamellites compared to that in the Ox Mountains metasediments as well as the discordance of the tectonic fabric with, and the lack of obvious reworking in, the metasediments is consistent with the hypothesis that the main focus of transcurrent deformation had shifted, possibly to the southeast (Chapter 6).

The sinistral fabric developed in both plutons is succeeded by discrete dextral shears and fractures that reflect minor late dextral deformation.

5.6 CONCLUSIONS : The relationship between kinematics, metamorphism and chronostratigraphy.

This is reviewed in figure 5.9 which illustrates that the inlier has experienced a complex structural history. Four main kinematic episodes are represented.

1. Initial fold and fabric development, and activity on the contact between the Cloonygowan Formation and the Ox Mountains Succession, (D2, Pre-478 Ma).

2. Sinistral transcurrent deformation, development of a braided fault system and syn-kinematic intrusion of the Ox Mountains granodiorite under amphibolite facies conditions at 478±12Ma (D3, Arenig-Llanvirn), followed by further sinistral deformation as the metamorphic grade decreased and the granodiorite cooled.

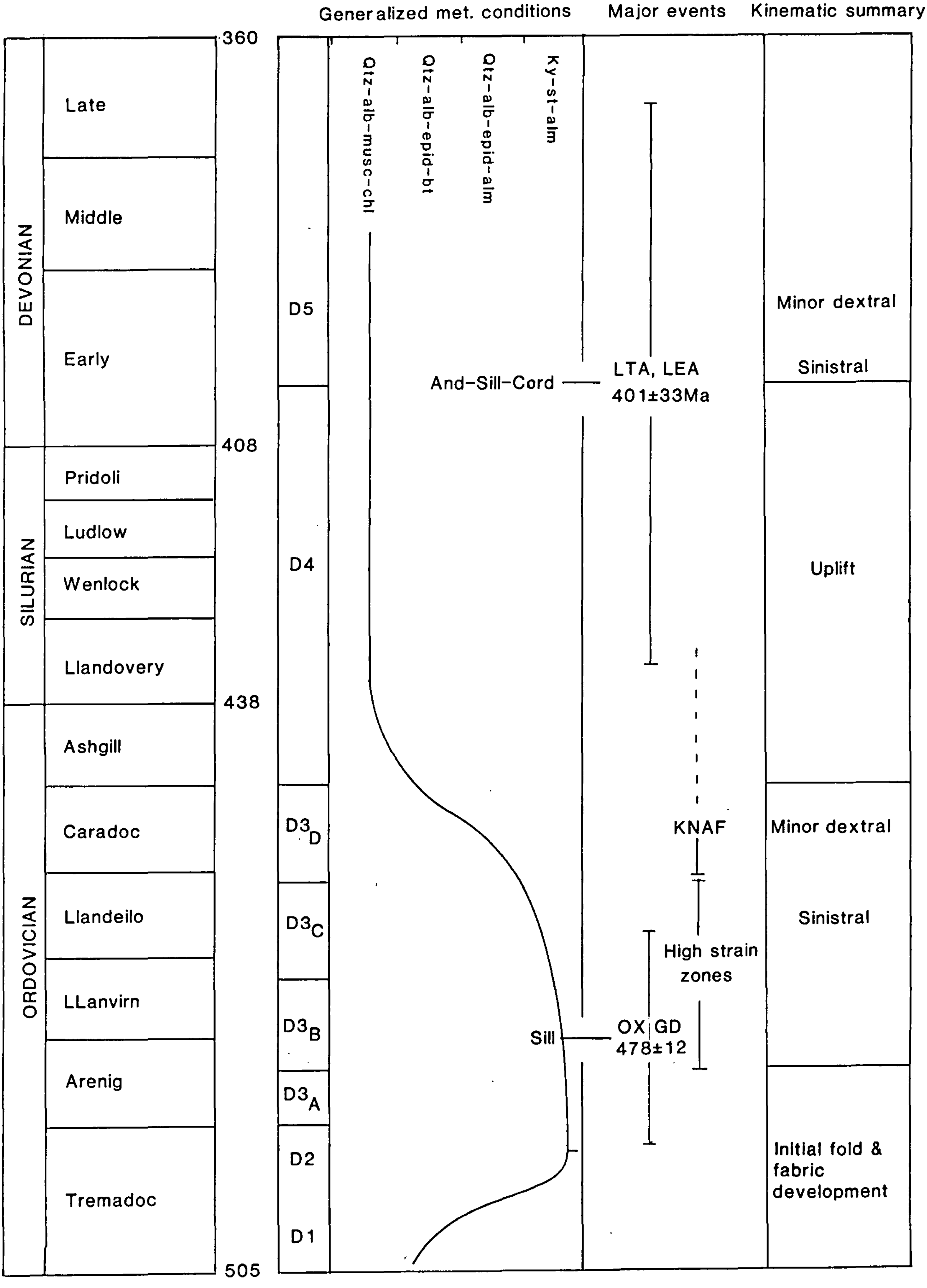
3. Uplift, reflected by decreasing metamorphic grade (D4).

4. Sinistral transcurrent deformation following intrusion of the Lough Talt Adamellite and Lough Easky Adamellite under greenschist facies conditions at 401±33Ma (D5, Early Devonian).

Figure 5.9 Generalized tectonothermal history of the Ox Mountains. Note: the intrusion of the Lough Easky and Lough Talt adamellites and the Ox Mountains Granodiorite provide the only chronostratigraphic markers for the deformation. The duration of each of the deformation phases is unknown.

FIG 5.9

GENERALIZED TECTONOTHERMAL MODEL



The rocks of the Cloonygowan Formation are considered to have experienced a similar structural history to those of the Ox Mountains Succession although at a higher structural level.

CHAPTER 6

REGIONAL TECTONIC SYNTHESIS AND CORRELATION

In this chapter, the position of the Ox Mountains in relation to the remainder of the Dalradian of the west of Ireland will be considered. The location, structure and kinematics of the Highland Boundary Fault Zone (HBFZ) will then be examined in detail for both the Irish and Scottish sectors. Evidence that this structure represents a major terrane boundary will be reviewed briefly, and the evidence for the existence of exotic terranes proposed by Winchester et al. (1988) in the Ox Mountains will be critically examined. The information that data from the Ox Mountains provides on the kinematics of the HBFZ and the constraints that this places on possible terrane accretion models for this structure will be discussed. Finally, moving further up scale and using a combination of the timing relationships of deformation to dated plutons and dated sedimentary rocks, an attempt will be made to correlate deformation in the Ox Mountains sector of the HBFZ with that observed along the remainder of the zone in Scotland.

6.1 INTRODUCTION TO THE CONCEPT OF TECTONOSTRATIGRAPHIC TERRANES

During the 1970s it became recognised that the Western Cordillera of North America are composed of a collage of fragments or *terranes*, each with a distinctive stratigraphy and separated by tectonic contacts (*tectonostratigraphic terranes*). Many terranes were identified that did not appear to have any relationship to the North American Craton, whilst others yielded evidence that they had been transported considerable distances. These displaced or *allochthonous* terranes that were proven to be far travelled became referred to as *exotic* (Coney et al. 1980; Dewey, 1982).

The term terrane was defined formally by Jones et al. (1983) as 'Fault bounded units of regional extent, each characterised by a geological history which is distinct from that of neighbouring terranes'. Terranes were therefore guilty until proven innocent : 'once the miogeocline or cratonic margin of an orogen is identified, all geologic terranes outside it are inherently suspect' (Williams and Hatcher , 1982). Recently, both the British and Irish Caledonides have been viewed as a collage of displaced terranes (Gibbons and Gayer, 1985; Hutton and Dewey, 1986; Hutton, 1987).

To apply this approach successfully it is essential to identify the continental foreland, miogeoclinal sediments or passive margin rocks. While shelf sediments of the Durness Group occur in the Northwest Highlands, the deeper water, passive margin sediments are apparently missing and the Durness sequence is separated from true oceanic rocks of the same age (Highland Border Complex) by Moine and Dalradian rocks. The deformation of the Moine and Dalradian is dominated by strike–perpendicular shortening; it is therefore extremely unlikely that strike–slip has removed the missing miogeoclinal sediments. Hutton (1987) stated that the Moine and Dalradian may represent at least the basement to the missing part of the Caledonian Miogeocline. If that is true; the major fault (The Highland Boundary Fault) which separates these rocks from Caledonian rocks of undoubted oceanic affinity is a major terrane boundary.

Hutton (1987) assigned the rocks of the Dalradian Supergroup that lie immediately to the north of the Highland Boundary fault and were subject to across strike shortening during the Grampian orogeny (500Ma), and uplift between 460 and 440Ma (Dewey and Pankhurst (1970), as the *Grampian Terrane.*

The rocks that are exposed to the south of the Highland Boundary Fault consist of a pre–Arenig ophiolite overlain unconformably by a shallow water Arenig Limestone, which is in turn overlain unconformably by a mid–Ordovician sequence of black shales, pillow lavas, cherts and agglomerates with clastic wedges that clearly represent a deep water sequence (Curry et al. 1982, 1984; Bluck et al. 1984). The recent work of Harper et al., (1989), and Bluck et al., (1984), has demonstrated that the rocks of the Highland Border complex, assigned to the Highland Border Terrane (Hutton 1987), that occur at Stonehaven, Arran and Aberfoyle in Scotland are correlatives of at least part of the Clew Bay Supergroup in Ireland. Thus the terrane, which never exceeds 30km in width, extends along discontinuous exposure for some 700km along strike.

Bluck (1985) demonstrated that the rocks of the Highland Border Complex are considerably younger than the adjacent Dalradian rocks. These Dalradian rocks were undergoing rapid cooling and uplift during the Ordovician (Dewey and Pankhurst, 1970), yet the Highland Border Complex lacks any Dalradian detritus. The Highland Boundary Fault is therefore regarded as a major terrane boundary during the Caledonian. Movement along this boundary must have been large and at least a minimum of the 700km strike length of the Highland Border Complex.

6.2 LOCATION OF THE HBFZ IN IRELAND

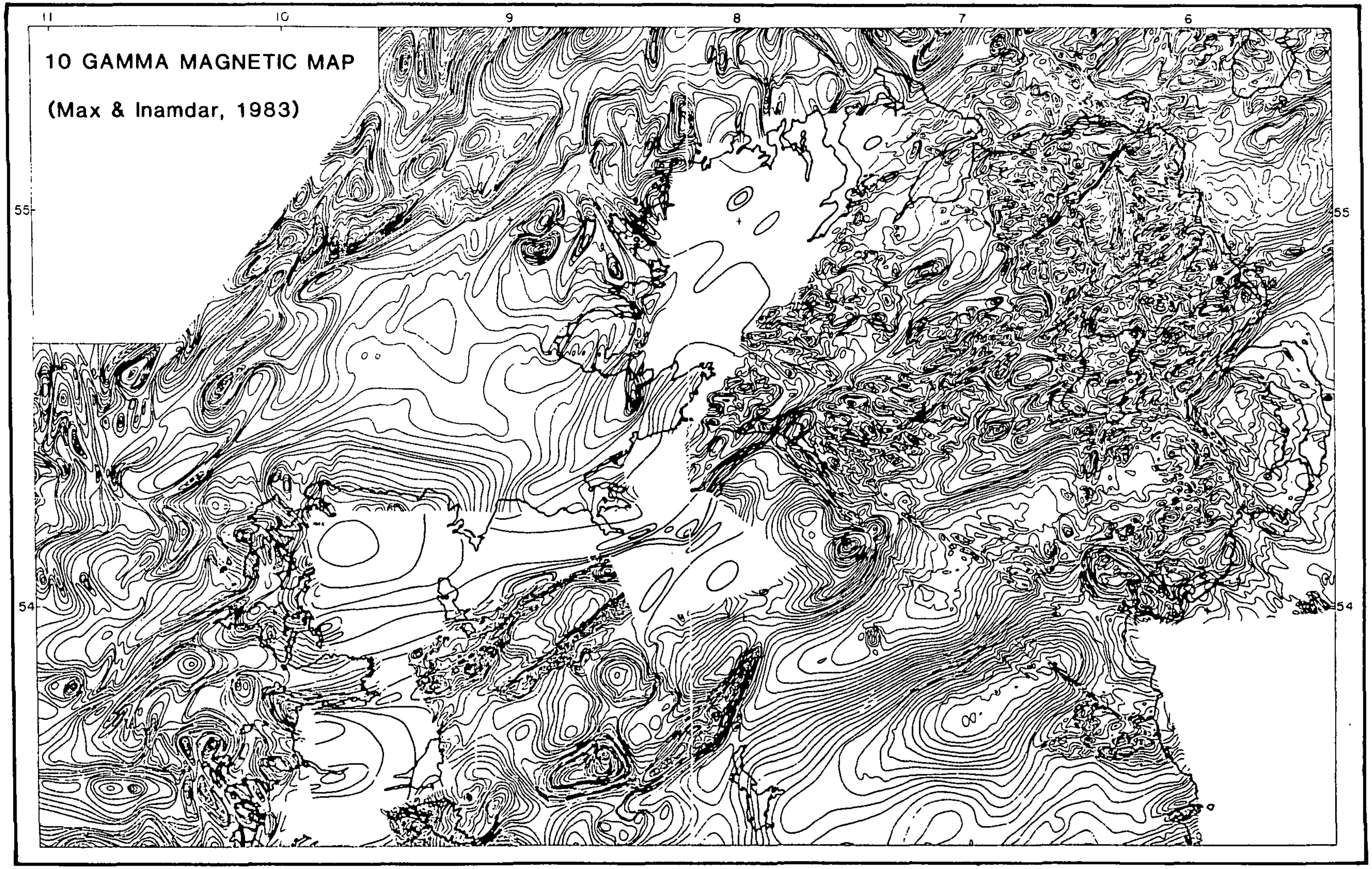
At its type localities on the British mainland, the Highland Boundary Fault, mainly affects Devonian rocks which give information only on its late kinematic history. The likely continuation of the fault in Ireland is the Fair Head – Clew Bay Line (FHCBL). Here the fault affects rocks of a wide variety of ages, which enables the early as well as the late kinematics to be studied. In this section the location of this structure is examined in detail before later discussing the kinematics of both the Irish and Scottish sectors.

Since Bailey and Holtedahl (1938) suggested the location of the Highland Boundary Fault in Ireland, some modifications have been made to follow more locally satisfactory geological lines, (Dunning et al. 1965; Dunning and Max 1975). Max et al. (1983) described the location of the Fair Head – Clew Bay line (FHCBL) as a primarily magnetic lineament and defined two well defined magnetic provinces: the northern Cratonic zone, consisting of largely negatively magnetized rocks with few important magnetic linears or bodies, and the southern segmented zone, consisting of low–moderate frequency, dominantly positively magnetic basement bodies. Max and Riddihough (1975) correlated this lineament with the Highland Boundary Fault in Scotland, and described it as separating a southern block, deformed in Middle to Late Cambrian time, from a northern block that was deformed in early Ordovician time. The data on which their interpretation is based is shown in Figure 6.1 and the interpretation itself illustrated in Figure 6.2. From these data a number of features are immediately apparent.

In Ireland the FHCBL is not always coincident with the local geological structure. Max and Riddihough (1975), state that the faults and serpentinite screens that occur along the south side of Clew Bay and on Clare Island (Phillips et al. 1969; Cole et al., 1914), previously regarded as the Highland Boundary Fault, occur entirely within the magnetically high, southern province and cannot be regarded as marking the line of the FHCBL. Max and Riddihough (1975) state that the magnetic signature is generally high across the Ox Mountains (they do not distinguish between the southwest, central and northeast Ox Mountains) and in the lower ground east of Clew Bay. They describe the north margin of the southern block as approximating to the 50 γ magnetic contour. This extends from the Achill Beg Fault in Clew Bay, runs along the northwest side of the Ox Mountains before trending east in the Lough Easky–Glennmore area of the central Ox Mountains, where it cuts across strike, transgressing the structure and stratigraphy of the Ox Mountains succession before intersecting the North Ox Mountains Slide at a high angle and running up the northwest side of the northeast Ox Mountains.

Max et al. (1983) recognised that the central and southwest Ox Mountains represent part of the Dalradian sedimentary succession and suggested that they have been thrust across the FHCBL in a manner similar to that observed in Tyrone. Hutton (1987).

FIG 6.1



10 GAMMA MAGNETIC MAP

(Max & Inamdar, 1983)

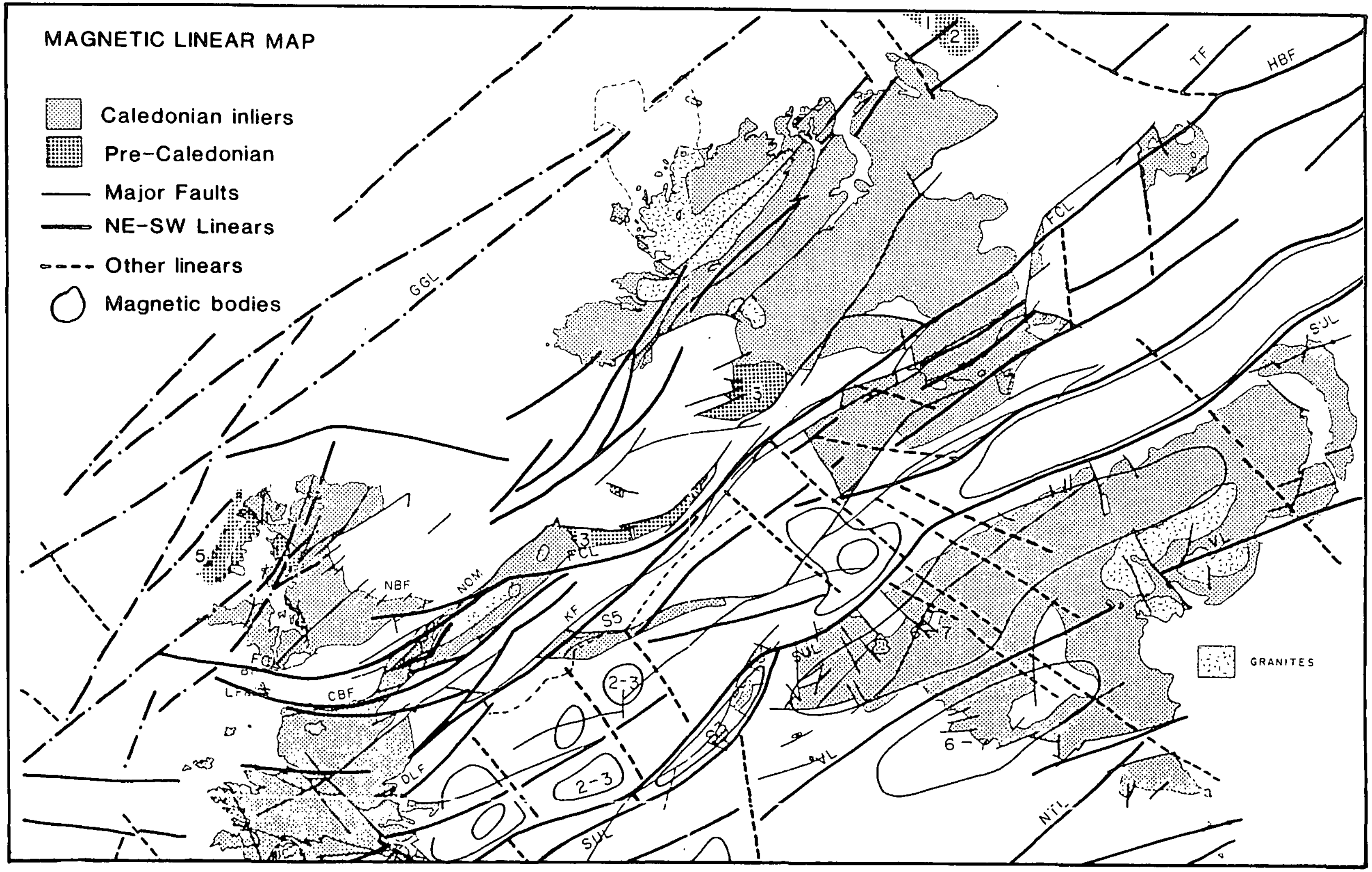

MAGNETIC LINEAR MAP
Caledonian inliers
Pre-Caledonian
Major Faults
NE-SW Linears
Other linears
Magnetic bodies
GRANITES
GGL
TF
HBF
FCL
SUL
LL
VL
NBF
NOM
FCL
KF
S5
FGL
LFA
CBF
DLF
2-3
2-3
SUL
NTL
1
2
3
5
5
6
7

FIG 6.2

This model necessitates a southeast translation of at least 10km. Synchronous with and following the intrusion of the Ox Mountains Granodiorite (syn D3), the tectonics of the Ox Mountains are dominated by sinistral strike–slip deformation. If this translation occurred, it must therefore have taken place prior to D3. The limited geometrical and kinematic data available for deformation which preceeded the intrusion of the Ox Mountains granodiorite, does not enable this model to be either disproved or confirmed (see below).

Figure 6.2 illustrates that the region south of the arbitrarily defined boundary of the northern magnetic province (the 50 γ contour) contains a large number of anastamosing magnetic linears, some of which are concealed beneath carboniferous cover. Among the most important of these are the linears corresponding to the Knockaskibole fault in the southwest Ox Mountains and the magnetic linears that run along the southeast side of the inlier which extend to the south side of Clew Bay and is extrapolated to the Leck Fault on Clare Island.

In summary, the direct correlation of the surface trace of the HBFZ in Ireland with the magnetic linear or FHCBL as mapped by Max and Riddiough (1975) is inconsistent with the surface geology, as it cuts across both structural and stratigraphic boundaries in the Ox Mountains Succession. Two alternative models may account for this:

1. The suggestion that the Dalradian rocks of the central and southwest Ox Mountains may have been thrust over the trace of the FHCBL (Max and Riddihough, 1975) is not considered to be likely by the present author for two reasons. Firstly, there is no structural evidence to suggest that such thrusting occurred during Lower ORS times as in Tyrone. Secondly, geological evidence indicates that the intensity of transcurrent deformation increases towards the southeast. This strongly suggests that the boundary lies to the southeast of the Ox Mountains and not within it.

2. The magnetic linear map of Max and Riddihough (1975), (Figure 6.2) reveals that the FHCBL is not simply a single magnetic linear but is represented by a complex anastamosing series of magnetic linears refered to here as the HBFZ. This strongly supports the view of the present author (Chapter 5) that the HBF is expressed at depth as a braided fault system in which many major faults or ductile shear zones enclose blocks of less disturbed zones. It is considered that the linear mapped by Max and Riddihough (1975) is simply one of this large number of magnetic linears that makes up the 'Highland Boundary Fault Zone' and that it is entirely contained within the Dalradian sequence of the Grampian Terrane, (Section 6.3).

6.3 POSITION OF THE OX MOUNTAINS WITHIN THE BRITISH CALEDONIDES: EVIDENCE FOR TERRANES WITHIN THE OX MOUNTAINS

The detailed stratigraphic correlation of chapter 3 strongly suggests that the medium–high grade rocks of the southwest and central Ox Mountains represent direct correlatives of the middle Dalradian of central and southern Donegal. It is therefore clear that they are part of the Dalradian miogeocline or Grampian Terrane. Their location immediately northwest of the Highland Boundary fault means that the Ox Mountains represent the northwest sidewall to the Highland Boundary Fault. They therefore record the deformation that occurred along this structure, which is regarded as the most important terrane boundary in the British Isles (Section 6.1), and provide critical evidence on the kinematics of movement on this structure which is not available in Scotland.

It is evident from the provenance studies discussed in section 6.1 that the rocks of the Highland Border Complex represent an allochthonous terrane that is presently juxtaposed against the Dalradian, and that the deep water continental margin sediments that originally lay to the southeast of the Ox Mountains have been removed at least 700km along strike. Before examining the kinematics of the Highland Boundary Fault Zone it is necessary to examine critically the evidence for the existence of additional terranes and terrane boundaries within the Ox Mountains that would further complicate the picture.

6.3.1 The Glennawoo Slide

In a recent paper, Winchester et al. (1987) presented geochemical data for Caledonised metavolcanic rocks from the Western Ox Mountains and the Northwest Mayo Inlier. They found geochemical differences between two metabasalt units (the Callow Member and the Newantrim Member) that occur within a thick metasedimentary sequence of semi–pelite and pelite in the Ox Mountains, and are separated by the Glennawoo Slide, (Section 4.3.5.2). These authors demonstrated that the metabasalts south of the slide have trace element compositions similar to mid–ocean ridge basalt (MORB), whereas metabasalts north of the slide have compositions similar to continental tholeiites of the ocean island basalt (OIB) association. They concluded from these data that the two groups of metabasalts were erupted in different crustal settings and proposed the Glennawoo Slide as a major terrane boundary which subsequently brought the two units together.

This conclusion was critically examined by Jones and Leat (1988) who argued that the psammitic nature of the metasedimentary rocks closely associated with both groups of metabasalts indicates that neither was erupted in a truely oceanic setting and that both groups of metabasalts cound have been erupted in the same continental margin environment. The structural, stratigraphic and metamorphic evidence which precludes the identification of the Glennawwoo Slide as a major terrane boundary as defined by Jones et al. (1983) is discussed below.

Geochemical evidence : A recent trend in basalt geochemistry has been the increasing awareness that the degree of partial melting of mantle has a major influence on both major and minor trace element composition of basaltic melts (Thompson 1987, and references therein). A large degree of partial melting (up to 25% to satisfy major element requirements) of asthenospheric mantle generates MORB which has low abudances of trace elements that are incompatible (eg. LREE) relative to trace elements that are weakly incompatable (eg. HREE, Y) in mantle peridotite. Small degrees partial melting (c. 1%) of the same athenospheric mantle generates melts with high abundances of strongly incompatible trace elements relative to weakly incompatible trace elements. Such magmas would be tholeiitic or alkaline OIB or continental basalts (Fitton and Dunlop, 1985). Thus both the MORB-like metabasalts south of the Glennawoo Slide and the LREE–enriched metabasalts north of the slide could have been derived from the same asthenospheric source mantle by different degrees of partial melting. In a rifted continental margin, basaltic volcanism is linked to lithospheric stretching (β>2; White et al. 1987). When the amount of stretching is large, and the degree of partial melting of the mantle is large, MORB magmas will be generated (Thompson,1987). When the amount of stretching is less and the degree of partial melting is small, LREE–enriched, tholeiitic or transitional to alkalic basalts will be generated, as erupted in many continental rifts.

The rate of stretching at a newly formed rifted continental margin will be different at different times. It follows that a range of asthenosphere–derived magma types from MORB to LREE–enriched tholeiite, and even alkaline basalts is expected on any section of a rifted continental margin. Morton and Taylor (1987) documented a strong case for MORB–like basalts having been erupted onto thin continental lithosphere on the margin of the North Atlantic. There is therefore no convincing reason why the metabasalts of both the Callow Formation and the Newantrim Member could not have been erupted onto the same continental margin (ie. in the same terrane). It is thus possible to explain the differences between the two chemical types of metabasalts by different degrees of melting of asthenosphere in an intraplate environment. There is no need to invoke magma generation in the subcontinental lithospheric mantle in this case (Leat pers comm, 1988).

Stratigraphic, structural and metamorphic evidence : The Ummoon and Lismoran Formations (sections 2.1.4, 2.1.3) were originally defined by Taylor (1969) who recognised that they are separated by a zone of intense strain which he named the Glennawoo Slide (Section 4.3.5.2). These Formations are very similar and both consist of psammite, semi–pelite and pelite. The present author agrees with Phillips et al. (1975) that the original transition between these units is still identifiable and that the present boundary between them is simply an arbitrary one, and does not represent a significant change in sedimentary environment or sediment supply. This precludes their identification as separate tectonostratigraphic terranes and demonstrates that the amount of movement on the Glennawoo Slide was unlikely to have been very large.

Detailed structural studies (Chapter 4) indicate that the rocks on both sides of the GWS have identical deformation histories D1–D4. The D3 deformation is represented by a steeply inclined penetrative fabric and upright folds whose axes are parallel to the well developed stretching lineation. These F3 folds progressively tighten approaching the Glennawoo Slide, indicating that it represents a local intensification of D3 strain, the gently plunging stretching lineation suggesting a dominantly strike–slip movement and kinematic indicators consistently indicate sinistral shear. Although the strains are locally high within this zone the present author considers them to be insufficient evidence for the minimum of 70km of strike–slip motion implied by the model of Winchester et al. (1987). Furthermore the identification of identical deformation histories on both sides of the Glennawoo Slide is inconsistent with the notion of two adjoining tectonostratigraphic terranes.

Winchester et al. (1987) contrast the grade of metamorphism on either side of the Slide as evidence of a terrane boundary. They show that the Newantrim Member contains the assemblage hornblende–oligoclase, indicative of the middle amphibolite facies, whereas the Callow Formation contains the assemblage hornblende–epidote–albite–biotite, indicating the slightly lower grade epidote amphibolite facies. Whilst it is true that such a variation could be attributed to juxtaposed allochthonous terranes, other factors may equally be responsible: for example, an original slight lateral geothermal gradient across the area, or a small vertical displacement component in the slide zone. Hence metamorphic arguments alone are inconclusive with respect to the 'terrane' hypothesis of Winchester et al. (1987).

In summary, the present author considers that the geochemical difference between the two metabasalt units documented by Winchester et al. can satisfactorily be explained by their generation by different degrees of partial melting of the same asthenospheric mantle. Furthermore since the Glennawoo Slide separates rocks whose structural and stratigraphic histories are indistinguishable, there is neither a structural nor stratigraphic basis for regarding the Glennawoo Slide as a major terrane boundary.

6.3.2 The Callow Shear Zone

The medium to high grade rocks of the Ox Mountains Succession are considered to be direct correlatives of the Dalradian of central and southern Donegal ie. part of the Dalradian Continental margin sequence. The position of the Cloonygowan formation is, however, less clear. Following from the argument presented by Williams and Hatcher (1982) that 'once the miogeocline or cratonic margin of an orogen is identified, all geologic terranes outside of it are inherently suspect', it is therefore necessary to examine the possibility that the Cloonygowan Formation (that is always in tectonic contact with the Ox Mountains Succession) may represent an exotic terrane.

While stratigraphic evidence suggests that the rocks of the Cloonygowan Formation may be equivalent to the Southern Highland Group, the absence of chronostratigraphic data means that their precise stratigraphic position is in reality uncertain. The clear difference in stratigraphy and metamorphism from the high grade rocks of the central and southwest Ox Mountains to the northwest implies that the Cloonygowan Formation represents a separate tectono–stratigraphic terrane. The similarity in D1–D4 ductile structure can be explained if the terrane originally lay at a higher structural level in a transcurrent shear zone but was subsequently brought down into its present position by a late strike–parallel fault (section 4.8.1.9). The rocks of the tectonically bounded Cloonygowan Formation therefore represent an allochthonous terrane *senso stricto.* The common deformation history of this terrane with the Ox Mountains Succession which they are presently in tectonic contact however, indicates that this terrane originated in close proximity to the rocks of the Ox Mountains Succession.

6.4 STRATIGRAPHIC RELATIONSHIPS AND KINEMATICS OF THE CLEW BAY-OX MOUNTAINS SECTOR OF THE HBFZ

It is clear from both geophysical and geological data that the wide zone of intense transcurrent deformation that affects the Ox Mountains extends into Clew Bay and further west for an indeterminate distance. In the synthesis that follows an attempt will be made to correlate the kinematic history of the Ox Mountains with that established by other workers in the Clew Bay and South Mayo areas. This correlation draws extensively on the work of Max (1989) and other workers who are acknowledged in the following text. The geology of the components that make up the west of Ireland will be reviewed briefly (Figure 6.3) then the kinematic histories of these units and the tectonic contacts which separate them will be examined (Figure 6.4).

6.4.1 Tectonostratigraphic units adjacent to the HBFZ in the Clew Bay Ox Mountains Sector

1. Ox Mountains Succession: represents part of the Dalradian Succession and is correlated with the Dalradian Succession of central and southern Donegal, (Chapter 3). It therefore represents part of the autochthonous continental margin sequence.

2. North Mayo Inlier: is considered by Long and Max (1977) to represent part of the Argyll (Middle) Dalradian Succession and also part of the autochthonous continental margin sequence.

3. The Deer Park 'Schists' (Phillips, 1973; Max and Long, 1977) occur as a series of fault slices along Clew Bay, and form the main component of the Kill Inlier on Clare Island. This unit may represent basement (Phillips et al., 1976) or part of another Lower Palaeozoic Terrane (Max and Long, 1985), and contains a serpentinite, a basic igneous suite and black shales and cherts. It forms the southern margin of Clew bay and is considered to have the chemical characteristics of an ophiolite (Ryan et al. 1983).

4. The Cloonygowan Formation (including the Westport Grit) is exposed in the east corner of Clew Bay and in the southwest and central Ox Mountains. Long and Max assign these rocks to the Southern Highland Group on the basis of lithostratigraphic similarity. A fauna has not yet been recovered from these rocks; it has therefore not been possible to establish their age and stratigraphic identity with confidence. As they are always in tectonic contact with surrounding units these rocks are assigned the status of a tectono–stratigraphic terrane.

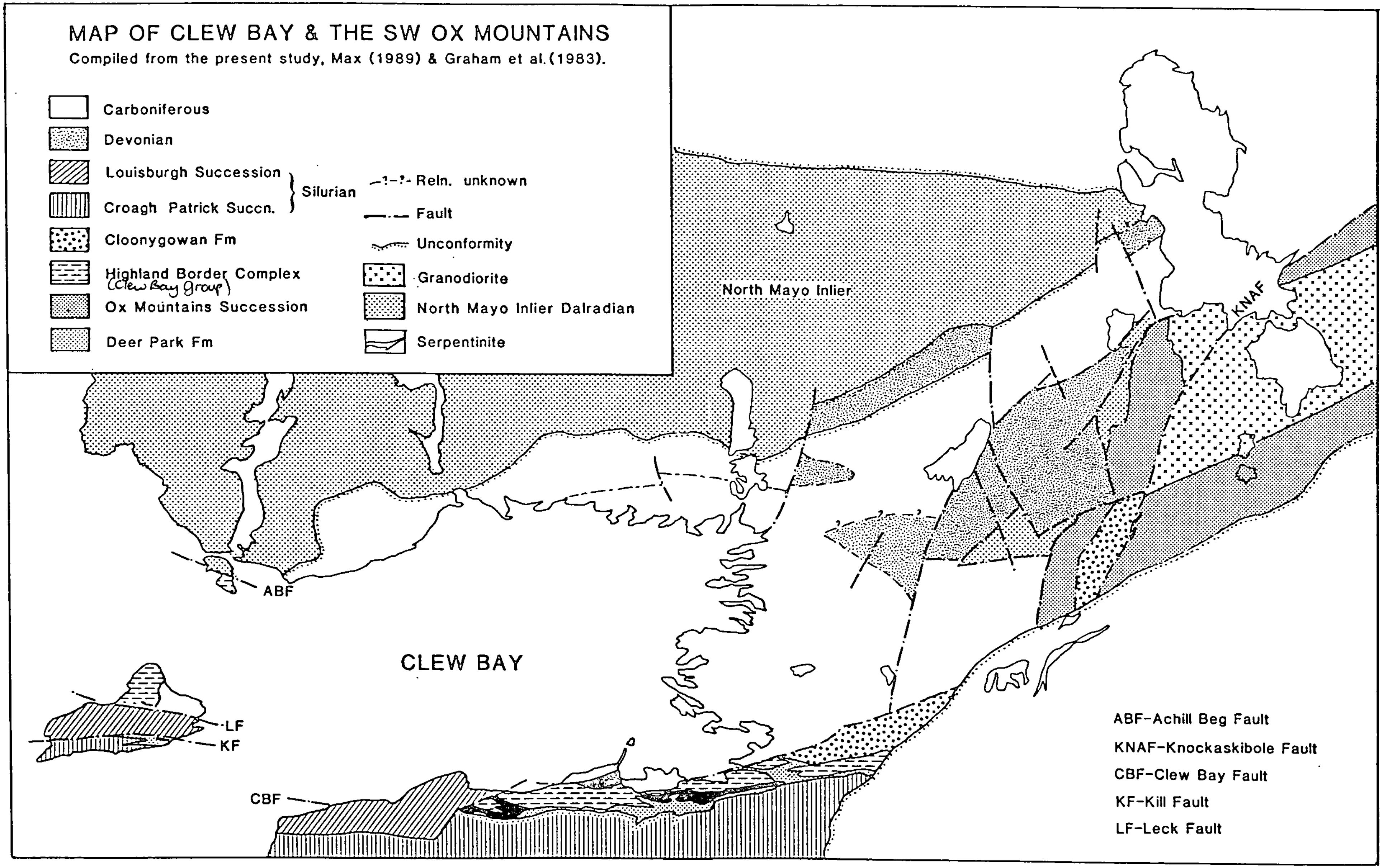

MAP OF CLEW BAY & THE SW OX MOUNTAINS
Compiled from the present study, Max (1989) & Graham et al.(1983).
Carboniferous
Devonian
Louisburgh Succession
Croagh Patrick Succn.
Silurian
Cloonygowan Fm
Highland Border Complex
(Clew Bay Group)
Ox Mountains Succession
Deer Park Fm
-?-?- Reln. unknown
Fault
Unconformity
Granodiorite
North Mayo Inlier Dalradian
Serpentinite
North Mayo Inlier
KNAF
ABF
CLEW BAY
LF
KF
CBF
ABF-Achill Beg Fault
KNAF-Knockaskibole Fault
CBF-Clew Bay Fault
KF-Kill Fault
LF-Leck Fault

FIG 6.3

5. The Clew Bay Group (Max, 1989), that comprises marine turbidites, fine grained basic volcanics, black mudstones and conglomerates, includes the Ballytoohy Formation of Clare Island (Phillips, 1973) and Killadangan Formation (Graham et al., 1985) of south Clew Bay. It is exposed on Bills rocks south Achill Beg and the northern part of Clare Island.

Rushton and Phillips (1973) described Protospongia spicules from the Ballytoohy Formation of Clare Island which they interpeted as upper Lower to Middle Cambrian in age. Microfossils of similar age were also described from the Killadangan Formation (Bliss, 1987). Harper et al. (1989) presented a detailed stratigraphic correlation of the Clew Bay Supergroup with the Highland Border Complex of Scotland and found microfossils which were consistent with an Ordovician (Tremadoc–Llanvirn) age. This date was supported by the close lithostratigraphic similarity with the more reliably dated Highland Border Complex of Scotland. Hutton (1987) and Max (1989) also consider that these low greenschist facies metasediments and the structurally underlying relict ocean floor all comprise part of the Highland Border Terrane that is exposed sporadically along the south side of the Highland Boundary Fault.

6. Ordovician (Tremadoc–Upper Llanvirn) rocks lie between Clew Bay and Connemara to the south. In the north part of the area, this thick sequence of turbidites, shales, sandstones and conglomerates with volcanic horizons, ranges in age from Arenig–Upper Llanvirn, while in the southeast a Tremadoc–Arenig basic–intermediate volcanic sequence was deposited (Dewey & Shackleton, 1984).

7. Silurian (upper LLandovery–Wenlock) sequences of the Croagh Patrick Succession (Bickle et al., 1972), the Clare Island–Louisburgh Succession and Killary sequence occur widely in the Clew Bay Graben. Llandovery–Wenlock rocks deposited in the footwall of the Clew Bay Fault rest unconformably upon sheared serpentinites and the Clew Bay Group and conceal an inferred tectonic contact with the Ordovician rocks in the northern South Mayo Trough.

FIG 6.4

		Ox Mountains	Clew Bay Group	SMT	Devonian	Achill Beg	Clare Island
					Maam Group		
360							
DEVONIAN	Late						
	Middle						
	Early	Minor Dextral SINISTRAL LTA LEA (401-33Ma)	Clew Bay Group thrust along Clew Bay Thrust over Silurian & SMT	Reactivation of the Clew Bay Thrust - Sinistral Minor dextral	Lower-Middle ORS deposited in a sinistral regime		
408							
SILURIAN	Pridoli			Croagh Patrick Syncline			
	Ludlow					Minor post Grampian Dextral	
	Wenlock	UPLIFT D4		Croagh Patrick Succession deposited on sheared serpentinites & Clew Bay Group Clare Island & Louisburgh Succession deposited between the ABF & CBF Killary sequence deposited across the Doon Fault			Minor Sinistral
	Llandovery						
438							
ORDOVICIAN	Ashgill						
	Caradoc			Rapid Uplift, folding & non deposition			
	Llandeilo						
	Llanvirn						
	Arenig	SINISTRAL OX GD (478-12Ma)	MOst sinistral transpression predates Silurian	SMT subsided		brings Upper Dalradian into contact with the Clew Bay Group	
	Tremadoc						
505							

8. Devonian, Lower–Middle ORS rocks were deposited in a sinistral extensional basin on the east side of Clew Bay (Graham, 1981b, 1983).

9. Undeformed Carboniferous strata unconformably overlie the Devonian and earlier strata.

6.4.2 Kinematics

The kinematic history of the Clew Bay - Ox mountains sector of the HBFZ is discussed below and summarized in Figure 6.4.

Ordovician : The earliest deformation within the Highland Boundary Fault Zone in Ireland is dated by the syn–kinematically deformed Ox Mountains Granodiorite at 478±12 Ma. This pluton was intruded into a dominantly transpressional environment and was deformed during its intrusion and crystallization. The major sinistral shear zone into which the Ox Mountains Granodiorite was intruded was at least the width of the outcrop of the Ox Mountains (approximately 10Km) and extended for an indeterminate distance west into Clew Bay and beyond. The sinistral transpressive deformation history of the steeply north–dipping Achill Beg Fault, which brings the Dalradian of the North Mayo Inlier into contact with the Clew Bay Group (Ryan et al. 1983), is therefore correlated with this early Ordovician deformation in the Ox Mountains. It is considered that faults present on the south side of Clew Bay (principally the Clew Bay Fault, since reactivated as the Clew Bay Thrust) are likely to have experienced a similar early history of sinistral movement. The duration of this early period of sinistral deformation is poorly constrained. It may have ceased as early as Llandeilo times, at which stage deformation in the South Mayo Trough had stopped and it was undergoing uplift and folding. If this were the case then the uplift which is widely recognised in the Dalradian at about this time (Dewey and Pankhurst, 1970: Richardson and Powell, 1976) and which occurs in the Ox Mountains (D4) may be correlated with this event. It is however possible that this period of intense sinistral transcurrent deformation could have continued for longer. The upper age limit is provided by the Upper Llandovery–Wenlock strata of the Croagh Patrick Succession which lie unconformably upon sheared serpentinites and the Clew Bay Group on the south side of Clew Bay. This age relationship is also supported by the Silurian Clare Island–Louisburgh Succession which was deposited between the Achill Beg and the Clew Bay Faults and which are only weakly deformed (Max, 1989) This indicates that the main components of the Highland Boundary Fault Zone in Ireland were assembled pre-Upper Llandovery times and thus overlapped with the amphibolite facies deformation of the Ox Mountains Granodiorite. The dominant tectonic pattern therefore existed pre-Upper Llandovery.

Silurian : The early Silurian (Llandovery–Wenlock) is a period of relative tectonic inactivity. It is dominated by the deposition of the red beds, shallow marine and turbidite sediments of the Louisburgh Succession, Croagh Patrick Succession, and Killary Sequence. The Llandovery–Wenlock Croagh Patrick Succession is deposited in the footwall of the Clew Bay Fault. These rocks record minor dextral deformation, which may be correlated with similar weak dextral fabrics recorded on the Achill Beg Fault and in the footwall to the Emlagh Thrust (J. D. Johnstone, pers comm). On the south side of Clew Bay sinistral deformation at chlorite grade is widespread in the Silurian rocks (this study and also see Hutton and Dewey, 1986). According to Johnstone this overprints the earlier minor dextral deformation. The amount of sinistral displacement is constrained to be small by metamorphic considerations: greenschist facies rocks are brought into contact with greenschist facies rocks, the maximum vertical displacement is therefore approximately 5Km, which, given the moderately plunging stretching lineation, constrains the maximum horizontal displacement to be approximately 10Km. Minor sinistral deformation is also recorded in the Silurian rocks on Clare Island. The timing of deformation in these Silurian rocks cannot be earlier than Wenlock. It therefore overlaps the error bars on the timing of emplacement of the Lough Easky and Lough Talt Adamellites, which record sinistral deformation synchronous with their emplacement and cooling. The end of Silurian deformation involving major displacements is marked by activity on the Clew Bay Thrust which displaces the Clew Bay Group south, over the Ordovician and Silurian (Max, 1989).

Devonian : The earliest deformation recorded in the Silurian strata occurred post–Wenlock (the age of the youngest deformed rocks) and pre–Siegennian (Graham and Smith, 1981). This deformation is represented by the Croagh Patrick Syncline, interpreted by Hutton and Dewey (1986) as a major upright D1 fold, forming one side of a ductile flower structure produced by a sinistral transpressional shear zone whose axis is located in Clew Bay. These workers interpret the succeeding D2 deformation, which also produced upright folds with steep cleavages, as resulting from sinistral shear whose locus was along the southern bounding faults of the South Mayo Trough. The observation that the locus of deformation moved south with time is consistent with data from the Ox Mountains, which record only a minor amount of sinistral deformation in the Lough Easky and Lough Talt Adamellites, and show no reactivation of earlier sinistral structures or formation of major folds such as the Croagh Patrick Syncline. The most recent deformation recorded in the region produced small sinistral movements and minor thrusting (Currall, 1963) in late Devonian – Lower Carboniferous times.

6.4.3 Summary (See Figure 6.4)

The main deformation recorded by the rocks in the west of Ireland sector of the Highland Boundary Fault Zone occurred pre–Silurian. The main fault–bounded components of the west of Ireland were therefore assembled close to the present position before the major uplift recorded during the Silurian of the South Mayo Trough, Ox Mountains and elsewhere. Renewed sinistral deformation occurred during the Lower Devonian when final adjustments were made to the final position of the main tectonostratigraphic components. However, judging by the state of deformation, displacements were relatively small compared to those in the Ordovician. The main locus of deformation was situated south of the South Mayo Trough at this time. Minor dextral deformation is recorded sporadically. Its development is not considered by the present author to be sufficiently intense to have major tectonic significance within the study area. It can as easily be explained by minor readjustment between blocks, within an overall dominantly sinistral strike slip system.

6.5 THE KINEMATICS OF THE HBFZ IN TYRONE

In Tyrone, the exposed southeast boundary of the Grampian Terrane is a 30° north dipping thrust which has carried Dalradian rocks southeast over the magnetic lineament that here marks the trace of the HBFZ (Hutton, 1987). In some areas, Upper Devonian or Lower Carboniferous rocks are cataclastically and penetratively deformed along the exposed fault by oblique dextral overthrusting. This, combined with the magnetic data, suggests that the Dalradian and overstepping Devonian has been thrust southeast over the transcurrent fault. Evidence of earlier Ordovician or end-Silurian transcurrent movements, such as characterizes the HBFZ further to the southwest, is not observed.

6.6 STRATIGRAPHIC RELATIONS AND KINEMATICS OF THE SCOTTISH SECTOR OF THE HBFZ

In the following section, the chronology of major structural and metamorphic events in the Dalradian Supergroup and the relations of the Dalradian and the Highland Border Complex (Harte et al. 1984) are summarized. This provides the basis for comparison with the structure and kinematics of the Irish sector of the Highland Boundary Fault Zone.

6.6.1 Chronology of major structural and metamorphic events in the Dalradian Supergroup

Harte et al. (1984) consider that a common structural sequence can be demonstrated for the Dalradian rocks of the Southern Highlands, and that four principal structural phases are represented. They note that the degree and number of phases increases from south to north, such that adjacent to the Highland Border only D1 structures are seen, and that these increase in intensity and the D2–D4 phases appear northwards towards the flat belt. Harte and Johnson (1969) have shown that sillimanite growth and migmatite formation in Angus represent the culmination of a single progressive metamorphic episode during or after D3. In the flat belt of Perthshire, the growth of porphyroblasts is pre-to syn-D3 (Dempster, 1983). There is therefore clear evidence in the Southern Highlands, that peak metamorphic conditions were attained close to D3 age. Retrograde metamorphism is widespread and apparently associated with the D4 structural event.

The Ben Vuirich granite, which was emplaced between D2 and D3 and therefore close to the peak of regional metamorphism has been dated at 514±15Ma using U–Pb from fractions of unabraided zircons (Pankhurst and Pidgeon,1976). This is consistent with the Rb–Sr wholerock age of 481±15Ma for the syn–D3 Dunfallandy Hill granite (Pankhurst and Pidgeon,1976), and close to the Rb–Sr age determination for the Ox Mountains Granodiorite.

A number of post–metamorphic granites have been dated in the northeast Grampians which give Rb–Sr wholerock ages in the range 470–440Ma (Pankhurst 1970, 1974). This is close to the K–Ar and Rb–Sr ages on hornblende and micas in Angus and Perthshire of 460–440Ma, that date the rapid uplift and cooling which affected extensive areas of the flat belt (Dempster 1983, 1984b). In the Southern Highlands the majority of the granitic intrusions now exposed were intruded during the final stages of the Caledonian orogenic activity at around 400Ma.

The recent work of Rogers et al., (1989) who dated the Ben Vuirich Granite by U–Pb analyses of abraided zircons at 590±2 Ma, has raised doubts about the absolute age of the British Caledonides. At the present time age determinations are also being revised for the Ox Mountains Granodiorite and other plutons. Until these age dates are known and the fundamental problem of the absolute age date of the Caledonian is resolved it is imprudent to base or indeed reject any structural correlation between areas on the basis of uncorroborated age dates from individual plutons, especially when these are derived by different analytical methods.

6.6.2 Evidence for the HBFZ in Scotland representing a major structural lineament

The parallelism of major D1 fold traces to the HBFZ suggests that this structure exerted a control on the nappe style and nappe generation in the southwest of the zone (Harte et al., 1984). The southern limit of significant D2 deformation in the steep belt has a striking parallelism with the downbend axis and the Highland Border. This line appears to be the intersection of a roughly planar boundary surface to D2 deformation with the present erosion surface. This line intersects the nappe boundaries and does not correspond to any particular level or depth. D3 structures are only occasionaly developed adjacent to the HBFZ. The occurrance of a well defined southern boundary to D3 deformation resembles the situation seen for D2 and once more suggests that the HBFZ was a major control on D3 deformation. In the Southern Highlands D4 produced a major monoclinal structure, the Highland Border downbend. The scale and continuity of this structure reflects the importance of the HBFZ during D4. The downbend implies a downstep of approximately 10km to the southeast. It therefore probably results from the reactivation of a subvertical fault within the basement.

6.6.3 Age and structural history of the Highland Border rocks

The youngest fossils found in the HBC rocks have a probable maximum age of Lower Caradoc (Curry et al. 1984), which means that the HBC rocks might all have been formed just prior to D4 in the Dalradian. Thus the timing of the juxtaposition of the Highland Border rocks against the Dalradian is constrained to be post–Caradoc. The Highland Border rocks must therefore have been in their present position relative to the Dalradian Supergroup prior to the unconformable deposition of the Late Silurian, Pridoli sediments.

On its southern side, the HBC is either faulted against the Devonian strata, or is unconformably overlain by Pridoli sediments. These relationships place upper constraints upon the timing of movement on the HBFZ. The contacts with Dalradian rocks to the north are tectonic, but their exact nature is often debatable; some contacts may have experienced several fault movements of different ages (Bluck 1984), but others may have been little disturbed since the initial emplacement of the HBC against the Dalradian Supergroup.

Harte et al. (1984) state that at Crichie Burn, 5km east of Glen Eske, the Dalradian and HBC metasediments are interleaved over a few cm and share a common ductile sub–mylonitic fabric. The Dalradian fabric becomes more intense approaching the contact with the development of a partly mylonitic fabric in the grits. There is however no unequivocal evidence for the age of the contact. Johnstone and Harris (1967) demonstrated similarities between the minor structural sequence of the HBC and Dalradian assemblages but could not see definite evidence for full correlation.

Harte et al. (1984) also note that there is no positive evidence favouring correlation of minor structures in the HBC and the Dalradian in Glen Eske prior to D4 in the Dalradian, and even at the D4 stage the evidence for correlation across the HBC units is not strong. Furthermore, a major problem exists in correlating early structures because of the absence of typical D1 and D2 structures from the Dalradian step belt adjacent to the HBFZ.

At Drumtochty the ORS is faulted directly against the Dalradian, obliterating evidence of earlier motions. At Garron Point, north of Stonehaven, the Lower ORS is unconformable on the HBC, the latter being in probable thrust contact with the Dalradian. The Lower ORS sequence is itself faulted. On Kintyre, and Arran the magnetic anomaly that marks the Highland Boundary fault, lies considerably north of the southern boundary of the Dalradian and the Lower ORS of Siegennian age oversteps the HBF, precluding extensive strike–slip movement later than earliest Devonian. Any major strike–slip motion must have occurred along lines within the Midland Valley now buried by younger sediments.

6.6.4 Emplacement of the HBC – Kinematics of the HBFZ in Scotland

Vertical displacements: The Dalradian and ORS rocks along the HBFZ commonly show steep attitudes which are related to the Highland Border Downbend and the Strathmore syncline. The magnitude of these displacements appears to decrease southwestwards, but they were clearly important during the formation of the downbend (460–440 Ma) through to the mid–Devonian.

Strike–slip displacement: If it is accepted that the HBC contains a major ophiolitic component, then its development must inevitably involve the juxtaposition of continental (Dalradian) and oceanic (HBC) crusts. The HBC ocean basin must have been remote from the Dalradian supergroup. Juxtaposition could have been achieved either by convergent (ie either subduction or obduction) or strike slip tectonics. Unfortunately in the Highland Border of Scotland, the minor structures that might be used to resolve this problem have not been reported. The emplacement of the HBC by strike slip movement is however favoured by analogy with the situation observed in the west of Ireland. Emplacement by strike slip tectonics easily accounts for the disappearance of the ocean basin in which the HBC rocks formed without requiring large amounts of compressive movement along the length of the Dalradian margin, late in its history and for which there is no supporting evidence. A relative vertical displacement may have accompanied this strike slip deformation and contributed to the emplacement of steeply orientated slivers of HBC rocks.

This raises the fundamental question of how the large component of transcurrent displacement demonstrated to occur along the Irish sector of the HBFZ is accommodated in Scotland ? Furthermore the absence of widespread evidence of intense sinistral strike slip within the Scottish Dalradian rocks adjacent to the HBC rocks and in the HBC rocks themselves, suggests two main possibilities. Either the mechanism of terrane accretion was different in the Scottish sector of the fault zone or subsequent brittle faulting has removed evidence for an early Lower Ordovician history of sinistral strike slip in the southern part of the Scottish Dalradian.

6.7 CORRELATION BETWEEN IRELAND AND SCOTLAND; KINEMATIC SUMMARY OF THE HBFZ (Figure 6.5)

Early Ordovician – Sinistral : The Ox Mountains and Clew Bay region both experienced an early period of intense sinistral deformation. The age of deformation in the Clew Bay region is constrained to be pre–Silurian, as Silurian sediments lie unconformably upon sheared serpentinites in the Clew Bay zone. The timing of deformation in the Ox Mountains is constrained by the synkinematic Ox Mountains Granodiorite, dated at 478 ± 12 Ma (although there is a verbal suggestion from the Geological Survey of Ireland that the pluton may be significantly younger). There is no small scale evidence for similar deformation along the Scottish sector of the HBFZ. Palaeontological data from the HBC in Scotland indicate that the HBC must have been emplaced post–Caradoc times, but before

FIG 6.5

			Ox Mountains	Clew Bay	Tyrone	Scotland
CARB		360				
DEVONIAN	Late				SE directed oblique dextral overthrusting	Carb overstep precludes motion in Scottish sector post L Carb
	Middle			Relatively undeformed M ORS basin constrains upper age limit of deformation (Graham, 1981)		Strathmore syncline pre-ORS
	Early		Minor dextral transcurrent defm. SINISTRAL (dated by-LTA LEA 401-33Ma	Clew Bay Gp thrust south over Ord & Sil of SMT		Arran : ORS oversteps HBF Drumtochty : ORS faulted against Dalradian Garron Pt : L ORS unconformable on HBC, HBC thrust over Dalradian.
SILURIAN	Pridoli	408		Main tectonic effects in Sil Sinistral Strike slip c 410Ma (Max 1989), displacement max 10Km Croagh Patrick Syncline		
	Ludlow					
	Wenlock					
	LLandovery					
ORDOVICIAN	Ashgill	438	UPLIFT D4	Most sinistral transpression predates Sil, which lies unconformably upon serpentinites & other rocks of the imbricate zone		Highland Border Downbend - upward movement to NW
	Caradoc					Earliest age for juxtaposition of HBC and Dalradian
	LLandeilo					
	LLanvirn					
	Arenig		SINISTRAL (dated by Ox Mts Gd 478-12Ma)			
	Tremadoc					
		505				

the deposition of an unconformable Silurian sequence. The duration of the early Ordovician period of sinistral deformation in the west of Ireland is unknown and may have continued until late Ordovician times. If that were the case then it is possible that the same event was responsible for the juxtaposition of the HBC and Dalradian in Scotland.

This initial period of sinistral deformation is succeeded by a protracted period of uplift between approximately 460 and 440 Ma, which is widely correlated throughout the Dalradian. There is also widespread evidence of retrogression at this time.

Silurian : The Silurian was a period of relative tectonic inactivity. Sinistral deformation is recorded in the Silurian rocks of the South Mayo Trough where the Croagh Patrick Syncline developed due to sinistral transpression during post–Wenlock to Siegennian times. Sinistral deformation of a similar age is also recorded in by the synkinematic emplacement of the Lough Easky and Lough Talt adamellites in the Ox Mountains, which are dated at 401 ± 33 Ma. The Clew Bay thrust is also reactivated as a sinistral structure at this time. This was the final major episode of transcurrent deformation recorded along the HBFZ. It is followed by the development of thrusts with relatively small displacements. In Ireland the Clew Bay Fault is reactivated as a thrust, along which the Clew Bay Group is carried over the Ordovician and Silurian of the South Mayo Trough. In Scotland the HBC was thrust over the Dalradian at a similar time. In Tyrone southeast directed, oblique overthrusting occurred slightly later, in the Late Devonian and displaces the southern margin of the Dalradian over the magnetic lineament that marks the location of the HBFZ.

The deposition of relatively undeformed Devonian sediments unconformably upon the Silurian places an upper limit on the age of deformation on the HBFZ.

CHAPTER 7

THE GEOMETRY AND KINEMATICS OF THE HIGHLAND BOUNDARY FAULT IN W. IRELAND: IMPLICATIONS FOR STRUCTURAL GEOMETRIES OF ACCRETIONARY TECTONICS.

Strike-slip faults are the key structures in the tectonic migration of terranes along plate boundaries. However Studies of the structural geometries and kinematics of accretionary tectonics and of terrane boundaries have been very limited. Howell et al. (1985) discuss the tectonostratigraphic terranes of the Circum-Pacific region and state that the identification of initial accretionary structures is extremely rare, and conclude that the mechanisms of terrane accretion and amalgamation remain obscure. Coney (1989) also acknowledges that the structural analysis of accretionary tectonics is a new field and by necessity must follow the regional stratigraphic and biostratigraphic studies necessary for terrane analysis. Most of the published accounts of terrane accretion in the Western North American Cordillera have lacked regional structural information. The study of the structure, geometry and kinematics of terrane accretion is therefore in its infancy, and a number of important problems remain.

1. What structural geometries and kinematics characterize the docking of terranes in strike–slip environments and, more generally, what is the geometry of strike–slip structures at deep crustal levels below the brittle–plastic transition ?

2. What is the role of basement in terrane accretion tectonics ? Does decoupling occur between basement and cover ? If so, at what level or levels ?

3. How are granitic intrusions introduced into shear zones and terrane boundaries such as the Ox Mountains ? Finally, is there a causal relationship between the

crustal thickening that accompanies transpressive deformation and the genesis of granitic melts, or do shear zones simply provide a favourable conduit for the ascent of granitic melts ?

The regional structural and tectonic review of chapter 6 demonstrates that the Highland Boundary Fault is a major terrane boundary and that the Ox Mountains is sited on the southeast margin of the Grampian terrane (Hutton, 1987), and represents the northwest sidewall of the Highland Boundary Fault. In addition evidence was presented that the intense and protracted history of strike-slip deformation recorded in the Ox Mountains is attributable to terrane accretion events along the Highland Boundary Fault. The Ox Mountains provides a unique view within the British Isles of the docking deformation recorded in the wall of a major terrane boundary (Chapter 6) and therefore provides new information to help develop solutions to the problems outlined previously.

7.1 THE STRUCTURE AND GEOMETRY OF TRANSCURRENT FAULT ZONES

It is necessary to understand the geometry of major transcurrent fault zones before examining in detail the role of basement in terrane accretion tectonics and the possible mechanisms of granitoid intrusion into strike–slip fault systems. There is an extensive literature on the geometry of strike–slip fault systems at or above the brittle–plastic transition. Much of this stems from the structural and seismic work carried out in relation to the hydrocarbon prospectivity of the upper levels of strike–slip systems. In this section this literature will be briefly reviewed, as it provides a useful basis for comparison for the study of the geometry of deeper level structures such as the Ox Mountains, which are less well documented.

7.1.1 High Level geometry; above the brittle–ductile transition.

Detailed mapping of strike–slip fault systems (Grose, 1959; Bridewell, 1975; Moore, 1979) shows that the upper levels of strike–slip faults are typically belts of braided, near vertical shears which bound elongate blocks. When strike–slip is convergent, blocks are squeezed upwards along these shears to make an upstanding source area for sediments. Where it is divergent crustal blocks may sag or tilt between bounding faults. Convergent strike–slip or transpression (Harland, 1971; Sanderson and Marchini, 1984) provides a component of horizontal shortening across the strike–slip fault zone which is necessarily accompanied by uplift of rocks in the fault zone. This is clearly demonstrated in laboratory models, where an elongate, fault bounded welt forms the zone of principal displacement because of the accommodation of the component of shortening strain by uplift (Lowell, 1972; Wilcox et al., 1973; Bartlett et al., 1981). The welt is bounded by sinuous faults which in cross–section are nearly vertical at depth and flatten upwards, carrying parts of the welt up on to the adjacent stable blocks. This upwards branching arrangement is mentioned by Willis (1938), Kingma (1958), Clayton (1966) and Sylvester (1988) and was mapped by Wallace (1949), Sylvester and Smith (1976), Burke (1979), Davies and Duebendofer (1982) and produced experimentally by Bartlett et al. (1981) and Naylor et al. (1986). Wilcox et al. (1973) termed this upward branching arrangement of faults a 'flower structure'. This upward flattening of strike–slip faults in zones of transpression has also been documented along the Alpine Fault in New Zealand (Wellman, 1955), along the Banning and San Jacinto faults of Southern California (Allen, 1957; Sharp, 1967) and in West Spitzbergen (Lowell, 1972; Kellogg, 1975; Craddock et al., 1985). Seismic profiles have been obtained over flower structures (Harding and Lowell, 1979; Bally, 1983; Harding, 1985).

7.1.2 Lower level geometry: the existence of a low angle detachment.

One of the principal problems is to understand what the geometry of strike–slip faults is at depths below that at which they appear to have a subvertical profile. Do they simply continue downwards through the lithosphere, maintaining their geometry observed at the surface ? Do they flatten off into a shallowly inclined detachment ? Or do they combine both geometries ?

The presence and structure of deeper shear boundaries and detachments is difficult to ellucidate because the seismogenic zone is confined to the upper continental crust in areas of active strike–slip faulting (Chen and Molnar, 1983; Sibson, 1983; Jackson, 1987). Therefore seismicity cannot give direct information on the location and nature of aseismic shear–zones and detachments that may exist in the lower crust and mantle. In this section the distribution of earthquake focal depths studied by Chen and Molnar, 1983 and the rheology profile through the lower crust , experimentally determined by Meissner and Wever, 1986 are reviewed as they provide important information on the likely location of detachment levels within the crust. Observational data on the geometry of normal, reverse and strike-slip faults is also reviewed as this provides a basis for comparison with exposed deep level strike-slip structures such as the Ox Mountains.

Chen and Molnar (1983) investigated the distribution of focal depths for earthquakes and found that the temperature of the source region was the most important factor in determining whether deformation occurs seismically or not. From estimates of the temperatures of the deepest events they concluded that for the crust the upper limit for brittle deformation is about 250–450°C and for the mantle this temperature is about 600–800°C. They inferred that both the upper crust and mantle seismic regions correspond to zones of relatively high strength, that are separated by a zone of lower strength in the lower crust, where ductile deformation predominates. Chen and Molnar concluded that this interpretation is in agreement with extrapolated values of experimentally determined brittle and ductile strengths of geological materials , (controlled by the flow laws of crustal and mantle materials) and suggested that a low strength zone in the lower crust might allow detachment of crystalline nappes from underlying mantle and lower crustal lithosphere (Figure 7.1a). They also pointed out that due to the uncertainty in the parameters used in the flow laws, the relative lack of knowledge of the mineralogy of the lower crust and possible roles played by water and other fluids, it is possible that the minimum strength occurs well above the Moho in the ductile crust.

Meissner and Wever (1986) compared the distribution of intracontinental seismicity in the upper crust with theoretically determined strength–depth curves (Figure 7.1b). They concluded that the similarity between the seismicity–depth curves and the strength–depth or viscosity–depth curves are a strong indication that the lower crust is a zone of high ductility sandwiched between brittle layers. Thus the rheology depth profiles which they drew for different areas have a dog–tooth profile. An important feature of these profiles is the abrupt change in strength that occurs as major boundaries are crossed. These abrupt rheology contrasts, not only at the brittle–ductile transition but also within the lower crust and at the Moho provide ideal sites for the localization of major decollements.

In the following section the geophysical and field evidence for the existence of these detachments predicted by Meissner and Wever will be examined.

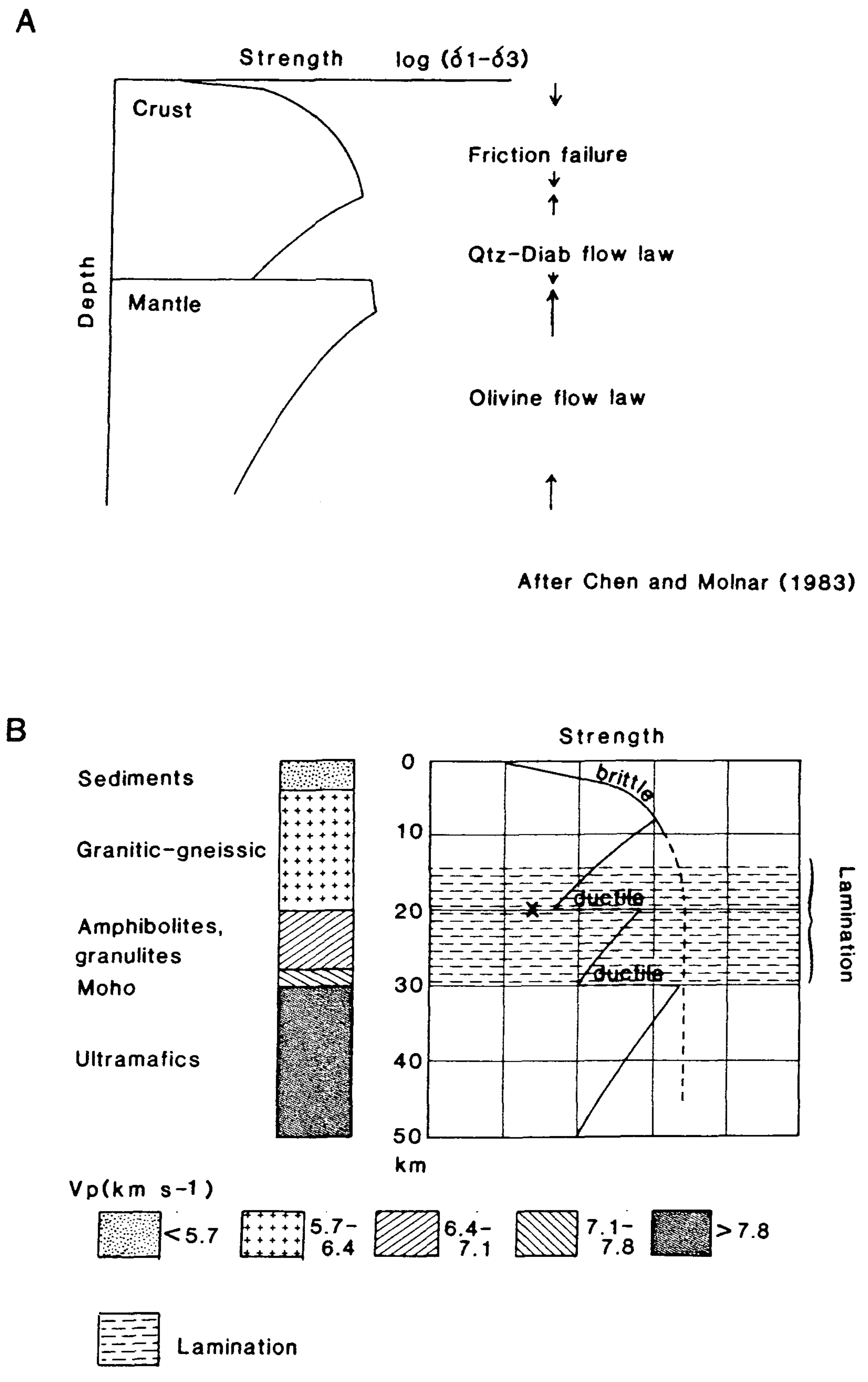
It is widely recognised that thrust faults involving crystalline basement commonly flatten with depth into decollement zones developed under greenschist or greater grades of metamorphism. In the middle or lower continental crust this forms partly or wholly crystalline thrust flakes which are characteristically around 10–15km in thickness (Oxburgh, 1972; Armstrong and Dick, 1974; Bally, 1981). Similar mid–crustal décollement regimes have also been recognised in association with extensive cover sequences strongly disrupted by listric–normal faulting (Wernicke and Burchfiel, 1982).

Figure 7.1

A. Schematic diagram of the variation of the mechanical strength of the crust and uppermost mantle with respect to depth. The upper part of the curve represents brittle failure and is based upon the linear relationship between the shear stress and normal stress or friction. The strength of the lower crust and upper mantle is controlled by the flow laws of crustal and mantle materials respectively, (After Chen and Molnar, 1983).

B. Simplfied velocity–depth column (for Variscan areas), and the strength–depth curve for the same crustal model. Vp, seismic velocity of compressional waves; Lamination represents the zone of strongly increased density of seismic reflectors (After Meissner and Wever, 1986).

Fig 7.1



After Chen and Molnar (1983)

After Meissner & Wever (1986)

Seismic evidence for thrust decollement at the base of the seismogenic zone comes from earthquake sequences in the vicinity of the 1971 M6.4 San Fernando earthquake in the Transverse Ranges of California, and from aftershocks of the 1978 M7.4 Taban earthquake in Iran. In both cases the main thrust ruptures dipped through the seismogenic zone at moderate angles of 50–55°, but focal mechanisms obtained from some of the deepest events were consistent with sub–horizontal shear surfaces (Hadley and Kanamori, 1978; Berberian, 1982).

Jackson (1987), states from work in the Aegean that 'not one single example of seismic activity has been recorded on a sub–horizontal fault' and concludes that faults with dips of <30° do not move seismically. However he acknowledges that it is possible that the steep upper crustal seismogenic faults flatten abruptly below the hypocentre into the lower crust where motion is aseismic, and states that the possibility of substantial aseismic motion on large extensional faults with dips of <30° cannot be excluded.

Jackson states that it is even less certain what happens in the lower crust. Several deep seismic reflection profiles have suggested the presence of discontinuities in the lower crust below the seismogenic zone (Brewer and Smythe, 1984). This is supported by recent work on actively extending areas such as the Gulf of Corinth (Jackson et al., 1982), where the asymmetry of vertical motions indicates that the fault that defines the southern margin of the Gulf projects beneath the northern bounding fault with a dip of approximately 10°. Jackson concludes that there is evidence from both vertical movements at the surface and seismic waveforms for very low angle faulting in the uppermost lower crust, beneath the depth at which earthquakes nucleate.

Thus there is evidence from thrust and extensional environments in support of a décollement horizon at the ductile–brittle transition as predicted from the data of Meissner and Wever. The situation for transcurrent fault zones is rather more complex. A series of possibilities (reviewed by Sibson, 1983) exist (Figure 7.2).

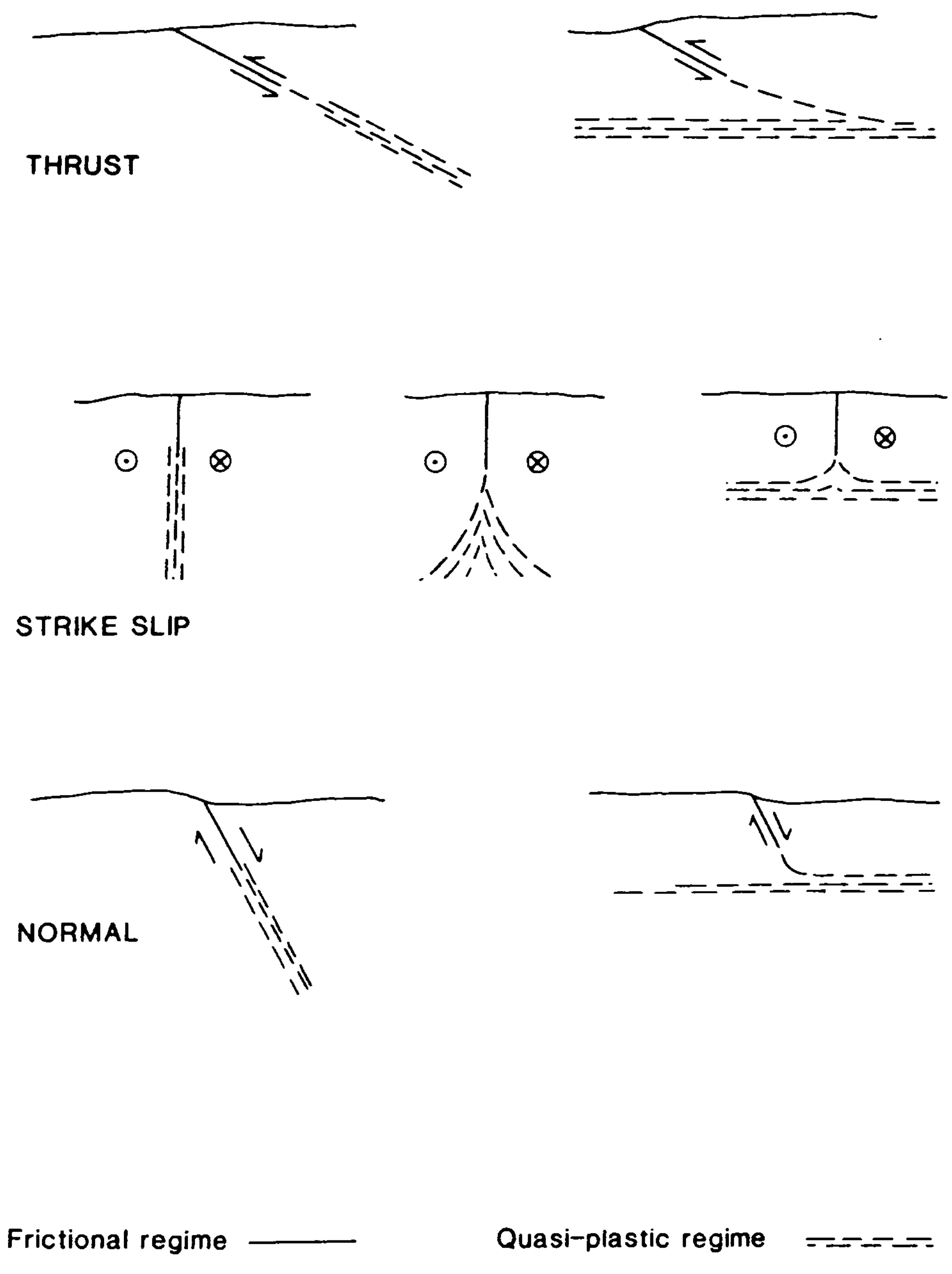
1. Deformation is concentrated in sub–vertical mylonite belts that continue downwards into quasi–plastic shear zones of constant width. Cogne et al. (1974), Juguso (1980), Gapais and Le Corre (1980) describe this relation from the Hercynian North and South Armorican Shear Zone which disrupts a crystalline basement complex, and provides a good example of a quasi–plastic, dominantly dextral strike–slip system formed in the middle crust under high greenschist–amphibolite facies conditions. In scale and complexity the whole transcurrent system of shear zones mimics the higher crustal level strike slip faulting patterns in S. California and New Zealand.

2. The Shear–zone widens with depth in the crust: Bak et al. (1975), and Sorenson (1983) described subvertical transcurrent gneissose shear–zones within crystalline basement in Greenland and demonstrated that the Nordre Stromfjord Shear–belt has a wedge shaped profile, decreasing dip and widening downwards with a wedge angle close to 40°.

3. An abrupt transition occurs between the steeply inclined upper element of the shear–zone and a sub–horizontal decoupling shear–zone: data predominantly from Western America infers the presence of a decollement surface or series of decollement surfaces. Palaeomagnetic analyses led Freund (1974) and Luyendyke et al. (1980, 1985) to suspect that parts of the American Pacific coast had rotated about a vertical axis in simple shear, and suggest that the rotated blocks must in fact be flakes (Oxburgh, 1972) or crustal panels (Dickinson, 1983). These detach on a shallow sub–horizontal shear surface at or near the base of the seismogenic zone thus giving a mechanism for regional rotation and translation of crustal slabs and flakes.

FIG 7.2

ALTERNATIVE MODELS FOR THE DOWNWARD CONTINUATION OF MAJOR FAULT ZONES After Sibson (1983)



Evidence that a mechanical discontinuity or detachment must underly the Transverse Ranges comes from the analysis of P–wave delays from earthquakes in southern California(Hadley and Kanamari, 1977; Nicholson et al. 1986) This is supported by deep reflection seismic profiling in the Mojave Desert and Transverse Ranges, which shows a series of nearly flat reflections in the upper crust (Cheadle et al., 1986). Additional P–wave analyses revealed that the distribution of relatively fast and slow crust is moderately well outlined by the position of the major faults in the upper crust, but not in the lower crust, suggesting that the major strike–slip faults in the Transverse Ranges do not extend below the seismogenic zone (Hearn and Clayton, 1986a, 1986b).

These studies have revived the notion of flake tectonics (Oxburgh 1972, 1974) with flake–like slabs up to 15km thick detached from the lower crust to offset strike–slip faults at depth (Hearn and Clayton, 1986a, 1986b; Dewey et al. 1986; Lemiszki and Brown, 1988).

Recent interpretation of seismic reflection studies of the intermediate and deep crust implies the presence of detachments perhaps at several levels in the crust, and that some strike–slip faults are cut at depth by those detachments. This work and the additional studies of Yeats, (1981), Terres and Sylvester (1981), Webb and Kanamori (1985), Nicholson et al. (1986) and Namson and Davis (1988) have raised many fundamental questions about the previously widely held assumption that all strike–slip faults are vertical to great depths. It is clear that in order to determine how general and widespread the development of these detachments is at lower levels additional observations, data and modelling is required. Although it is at an early stage at present this work does however provide a useful basis for comparison with exposed deep level strike–slip structures such as the Ox Mountains.

7.1.3 Data from the Ox Mountains

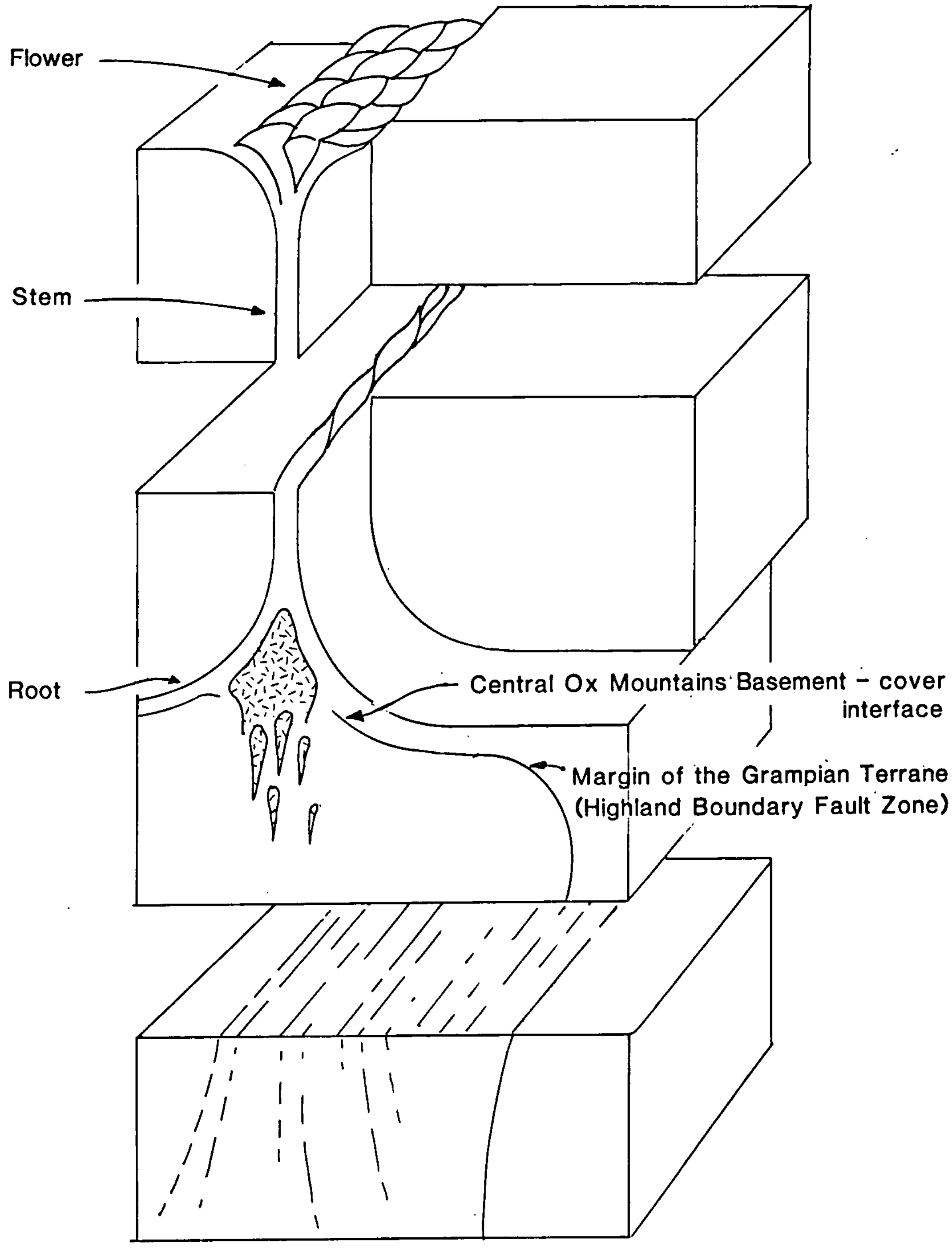
Data from the Ox Mountains (Chapters 4 and 5) suggests that at middle to lower crustal levels major transcurrent shear–zones may be expressed as an anastamosing series of high strain zones, that separate areas of lower tectonic strain. The geometry of these high strain zones indicates that the shear–zone broadens with depth. This geometry and in particular that of the flat–lying basement–cover interface exposed in the northeast of the area is consistent with this representing a shallowly inclined decollement surface. Elsewhere in the Ox Mountains the position is less clear. The absence of geophysical data means that the depth to basement and the geometry of the basement–cover interface is very poorly constrained in the remainder of the inlier to the southwest. An assessment of the likely geometry of the basement–cover interface and the probability of decoupling occuring between the basement and cover in the remainder of the inlier can, therefore, only be made by examining the geometry of the deformation within the cover sequence.

The geometry of the anastamosing series of high strain zones that make up the inlier (figs 5.5 - 5.8) is consistent with the shear–zone diverging at depth, which suggests that the structures may shallow out into a basement décollement and have the geometry of a 'root' structure, (Figure 7.3). This contact represents a significant rheology contrast between the largely anhydrous granulite facies psammites of the northeast Ox Mountains and the overlying metasediments of the Ox Mountains Succession which contain up to 50% phyllosilicate. Analogy with the data of Meissner and Wever (1986), (Figure 7.1b) suggests that this contact may represent a major decollement horizon produced by a rheology change such as X of Figure 7.1b

This does not however preclude a strand or strands of the system extending deeper in the crust to form a conduit along which magma can be easily introduced into the system.

Figure 7.3 A highly schematic block diagram illustrating the deep and shallow level geometry of a major transpressional fault system such as the Ox Mountains.

FIG 7.3


Flower
Stem
Root
Central Ox Mountains Basement – cover interface
Margin of the Grampian Terrane (Highland Boundary Fault Zone)

All of these data and information gained from other areas has been integrated to produce the highly schematic model (Figure 7.3) for the geometry of transpressive strike–slip systems at high and low levels in the crust. This model predicts that low angle detachments exist at a number of levels,

1. At high levels, above the exposure level of the Ox Mountains, where they will be concentrated at the base of the seismogenic crust and,

2. At lower levels, in the case of the Ox Mountains at the level of the basement–cover interface, which is a significant rheological boundary.

3. In addition, the possibility of further detachments at even deeper levels, as suggested by Chen and Molnar (1983), cannot be excluded.

The resultant geometry of the transpressive strike–slip system is therefore one in which a series of high strain zones anastamose both horizontally and vertically producing a series of lozenge shaped areas bounded by areas of high strain. The existence of a strongly developed and steeply inclined fabric and therefore a strong planar anisotropy may facilitate introduction of magma into the strike slip system and strongly influence the emplacement geometry.

7.2 THE ROLE OF BASEMENT IN TERRANE ACCRETION TECTONICS IN TRANSPRESSIVE STRIKE–SLIP ENVIRONMENTS

One of the problems that must be addressed in any terrane accretion model is to understand the role of basement in accretion mechanisms. Are terranes essentially thin–skinned sheets that are detached from their underlying basement ? Is basement accreted along with the higher level cover ? Finally, does the involvement of basement change depending on the regional tectonic environment ?

Terrane accretion, as the term is applied along the west coast of America, assumes that, to some degree, oceanic lithosphere, which originally lay between a terrane and a continent, has been disposed of in some way, usually by subduction. Coney and Jones (1985) and Coney (1989) recognised that the distal terranes or oceanic terranes of North America are frequently found as thin thrust sheets sitting upon more inboard terranes and / or the miogeocline itself, and that the regional structural vergence is towards the craton. Coney states that the majority of these terranes seem to be composed of only supercrustal sedimentary cover and / or volcanic veneer, and that evidence for the presence of deeper crustal material is generally lacking, particularly in the more oceanic terranes and therefore concludes that the terranes represent thin slices or flakes, somehow detached from their original lower crustal and lithospheric basement. This leads to the problem of how this basement and lithospheric material were disposed of during the accretion process.

Price (1986) studied the Canadian Cordillera, where he concluded that the miogeoclinal (or ancient continental margin) and parts of the platform cover and foreland basin deposits have become detached from their basement and displaced over the flank of the craton. This material is therefore bounded above and below by shear zones or thrust faults. Price regards the mechanism responsible for this process as analogous to that which produces delamination in experimentally deformed anisotropic rocks which have been subjected to compression supplied sub–parallel to the layering and draws analogy between this process and the flake tectonics described by Oxburgh (1972, 1974) (Figure 7.4A) and Armstrong and Dick (1974). This model was developed to account for the removal of basement and lithosphere in enviroments dominated by orthogonal or near orthogonal convergence, and may not be directly applicable to environments such as the British Isles in which terrane accretion is likely to be dominated by strike–slip tectonics. In the discussion that follows three additional models of basement involvement in terrane accretion tectonics in a strike–slip environment are examined critically.

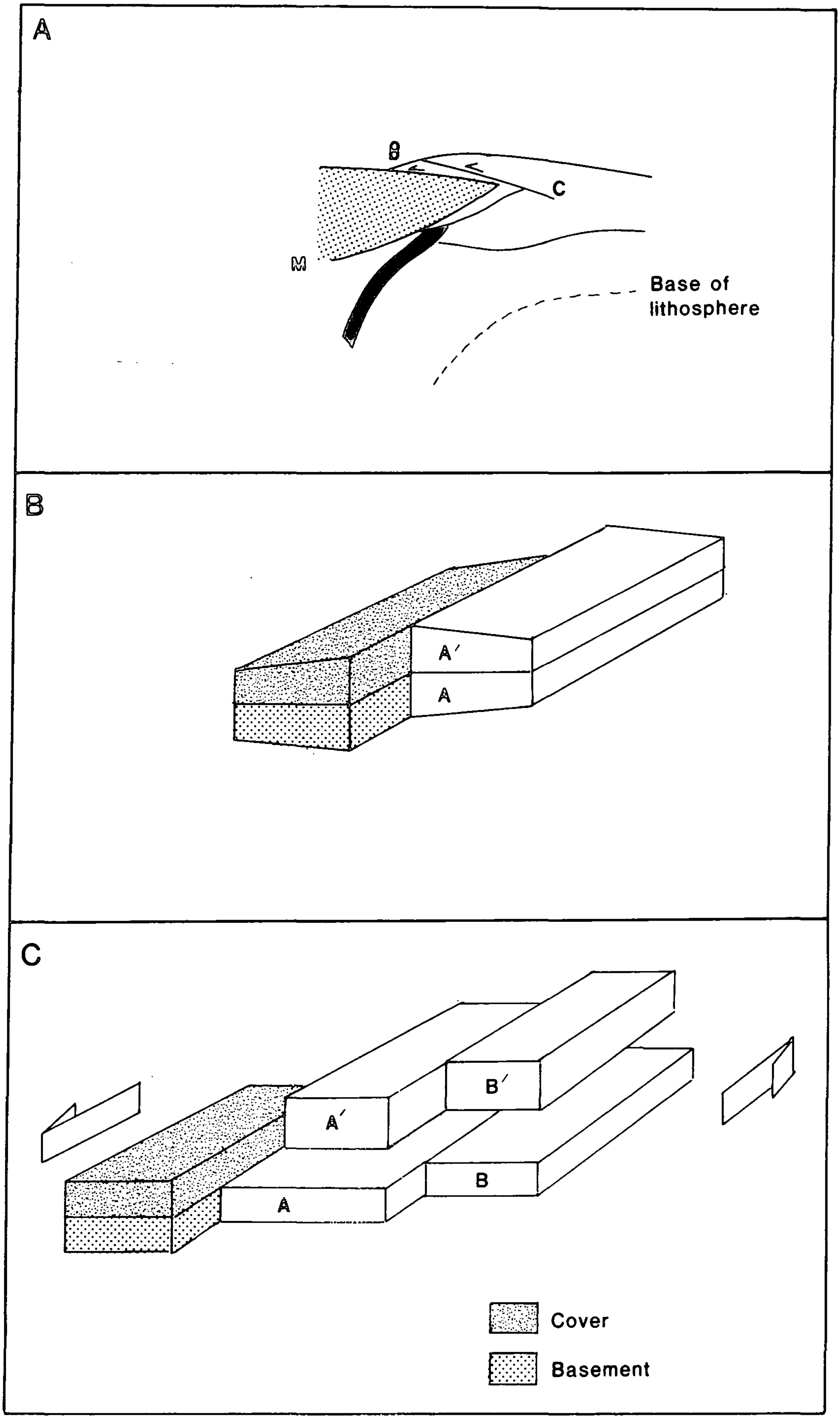
Figure 7.4 Alternative models for terrane accretion.

A. The flake tectonics model (Oxburgh, 1972). Gross geological relationships in the eastern Alps. A pre–Mesozoic basement complex A has been overridden by an allochthonous thrust mass of pre–Upper Palaeozoic metamorphic rocks C which is less than half the thickness of normal continental crust. This precludes the possibility that C can be a full plate in the usual sense of the word. It is therefore considered to be a thin upper crustal flake that has been detached from the southern plate. M, Moho; Black shadding, subducted oceanic crust; B, Perifer Schieferhulle; C, Altristallin Decke.

B. Terrane accretion where basement and cover remain attached during the accretion process, results in thickening of the basement and Lower crust and corresponding uplift at higher levels in the structure.

C. Terrane accretion where basement is detached from its original cover seqence may permit thin skinned terrane accretion. In this model the upper limit on the amount of accretion that can occur will be reached when accumulation of basement blocks A' and B' 'choke' the system.

FIG 7.4



1. If cover (A') and basement (A) (Figure 7.4B) remain attached during accretion and subsequent deformation of an allochthonous terrane in a transpressive environment, the docking deformation will be accommodated by thickening of the basement and lower crust and corresponding uplift at high levels of the structure. In this model basement and cover will be deformed together and basement will be accreted together with the overlying cover sequence.

2. Where cover is detached from its basement sequence along a shallowly inclined decollement thin–skinned terrane accretion can occur and allochthonous terranes composed entirely of cover metasediments and igneous rocks may be accreted onto the continental margin without their basement. In this model an upper limit on the amount of accretion that can take place will be reached when accumulation of basement blocks (A and B in figure 7.4 C), outboard of the continental margin 'choke' the system.

3. The final model (Figure 7.5) involves a combination of the development of a basement décollement and a braided strike–slip fault system. This results in the concept of basement 'stripping' where elongate slivers of basement material are removed by strike–slip motion along the axis of a braided strike–slip fault system. The geometry of the Ox Moutains shear zone illustrates that at deep levels in the crust transpressive deformation may occur by activity on an complex braided system of strike–slip faults that anastamose both laterally and vertically. This observed geometry is consistent with that outlined (Figure 7.5), and provides a viable mechanism for the shuffling and rotation of terranes, which cannot adequately be achieved by either of the first two models. The effect of this stripping of cover from its underlying basement will be to generate a series of cover terranes and also a series of basement terranes that are now remote from their cover sequences.

Figure 7.5 Basement stripping and the generation af basement and cover terranes.

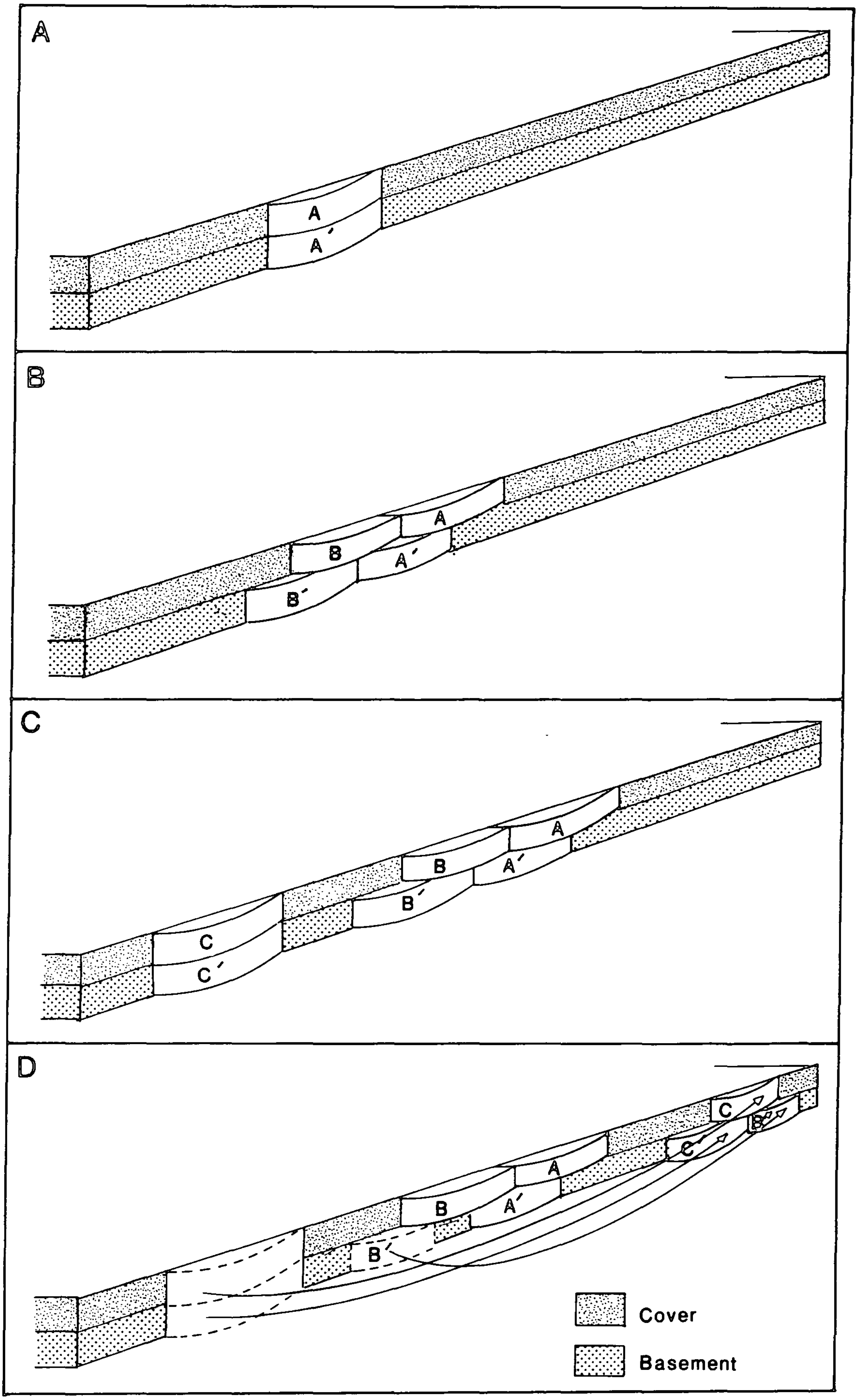
A. Accretion of terrane A against the continental margin by sinistral strike slip.

B. Accretion of additional terrane B, continuing strike slip deformation results in detachment of cover of terranes A and B producing basement terranes A' and B'.

C. Accretion of terrane C. A, A' and B, B' continue to be affected by sinistral strike slip.

D. Further activity on the sinistral strike slip system detaches terrane C and part of basement terrane B' (thus generating basement terrane B'') and transports them along the continental margin. This results in shuffleing of terranes A, B and C and in the isolation of basement terrane B' from its cover B.

FIG 7.5



7.3 COMMENTS ON THE GENESIS AND EMPLACEMENT OF GRANITIC INTRUSIONS IN MAJOR MID–LOWER CRUSTAL SHEAR ZONES

The association of granitic rocks with shear zones has been widely recognised (Pitcher and Bussell, 1977; Brun and Pons, 1981; Hutton, 1988). Much of the current research has sought to address two main problems.

1. Why does this association exist ? Is there a causal link between transcurrent deformation and granite genesis ? Or do shear zones simply provide a favourable conduit for the ascent of available magma ?

2. What emplacement mechanisms characterise the intrusion of granitic rocks in transpressional environments ?

The purpose of the discussion that follows, is not to review the extensive literature on granite genesis and emplacement (see Castro, 1987; Pitcher, 1987; Hutton, 1988 and references therein), but to comment on some observations made during the present study of the Ox Mountains that may be revelant to these problems.

7.3.1 The genesis of late–post tectonic granites: the Houseman, McKenzie and Molnar model.

When crust thickens during crustal shortening, the underlying mantle lithosphere must shorten and thicken also, causing cold dense material to submerge into the surrounding asthenosphere. Houseman et al.(1981) showed, for a range of physical parameters that the thickened boundary layer that forms the transition from the strong lithosphere to the convecting asthenosphere may become unstable and sink into the asthenosphere and be replaced by hotter asthenospheric material which then transfers heat to the crust and mantle above. These workers modelled this process using typical values for the physical parameters in the earth, and concluded that the boundary layer is removed over a period that is shorter than the duration of deformation in some collision zones (30–50m.y.). They suspected that in some cases the entire mantle lithosphere may detach from the lower crust during shortening, exposing the crust to athenospheric temperatures.

The thickening process will stretch the isotherm vertically in both the crust and uppermost mantle, reducing the geothermal gradient (Figure 7.6). If heat is applied to the base of the lithosphere at the same rate as before the shortening, then the geotherm should return to its original form. If however the re–establishment of the geotherm is fast in relation to the rate of erosion of the mountains formed by isostatically controlled uplift, then because the crust is anomalously thick, the lower crust and the top of the mantle will be atypically hot. Houseman et al. state that this may have two consequences:

1. The lower crust may melt and undergo very high temperature metamorphism, resulting in the production of regional metamorphism and the generation of late– to post–tectonic granites, after the major deformation has stopped.

2. Because olivine is stronger than most prevalent minerals in the crust, regions with relatively high temperatures at the Moho will be deprived by detachment of a layer of cold strong mantle. Thus if the geotherm is large ie. a region of shortening returns to normal before erosion eliminates the excess crustal thickness, the area is likely to remain weak and susceptable to further deformation.

7.3.2 The association of granites with shear zones

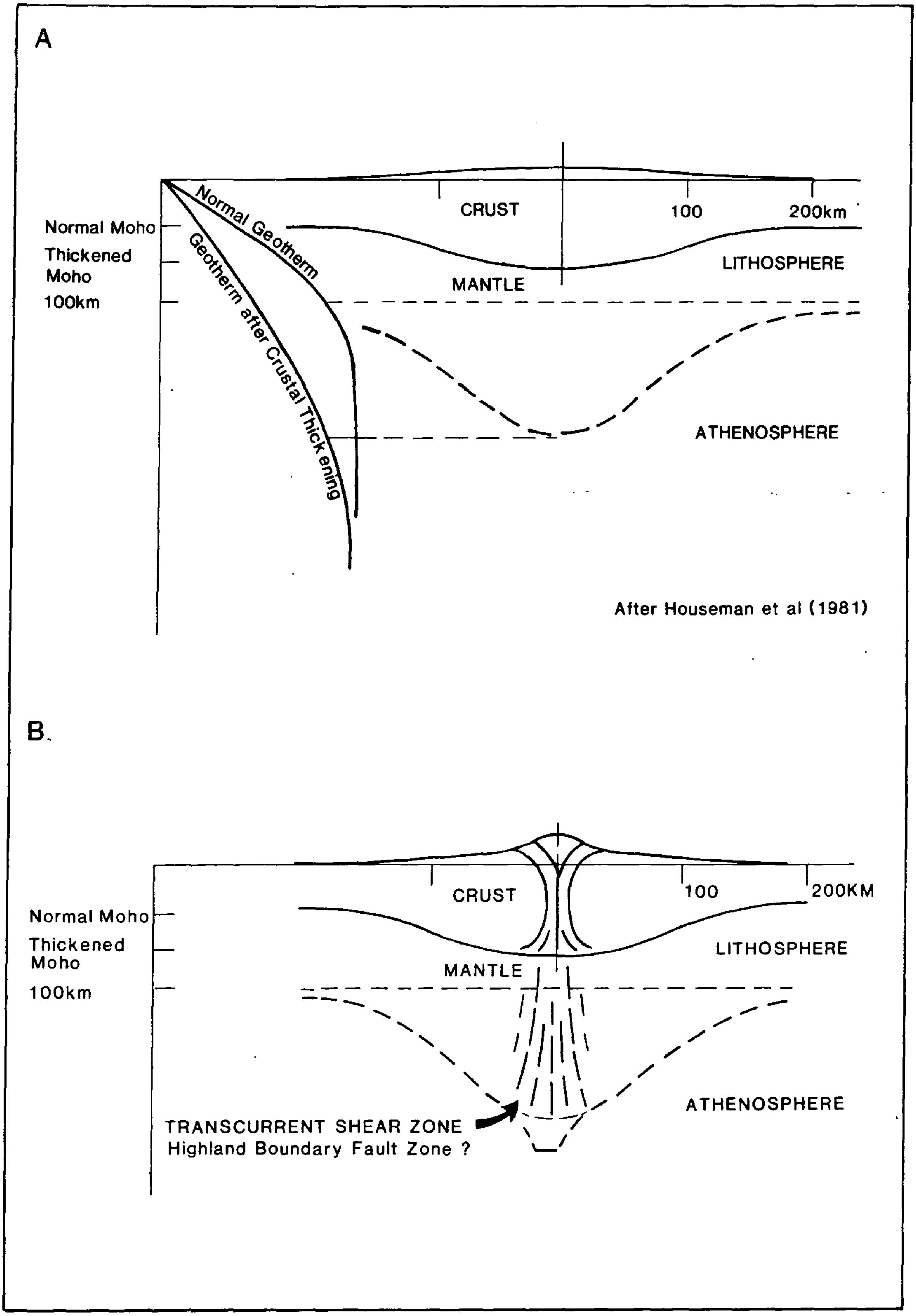
The widely observed association of granites with shear zones may reflect two features of shear zones. Shear zones may simply provide a favourable conduit for the ascent of magma or the genesis of granitic melts may be causally related to shear zone development. It is difficult to differentiate between these two

Figure 7.6

A. Simple cross section and geotherms to show typical scales in crustal thickening. In cross section the entire lithosphere is thickened by a factor of 2, and the original geotherm is stretched vertically by this amount. (After Houseman et al., 1981)

B. Effects of a major transpressional shear zone superimposed upon A, resulting in locally increased crustal thickening.

FIG 7.6



After Houseman et al (1981)

possibilities as both share the same result. In the following section the possible contribution of both of these factors will be examined briefly.

1. The geometry of granitoid intrusions in shear zones such as the Main Donegal Shear Zone and the Ox Mountains Shear Zone clearly illustrate that the pre–existing shear zone strongly influenced the geometry of the intrusion. The study of the geometry of the Ox Mountains Shear Zone (Chapters 4 & 5) demonstrates that, at mid–crustal levels, shear zones may be expressed as a number of high strain zones which anastamose both laterally and vertically. Thus shear zones represent a clearly defined planar anistropy within the crust which may provide a favourable conduit for the ascent of magma. Water exerts a fundamental control on the melt viscosity of granitic systems (Shaw, 1977; 1980). The presence of water within shear zones may therefore trigger the ascent of magmas to higher levels by reducing the effective melt viscosity.

2. Shear zones such as the Ox Mountains Shear Zone and the Main Donegal Shear Zone represent localized zones of transpression. The high transpressive strains associated with the Ox Mountains Shear Zone and the thrusting of metasediments of the southwest and central Ox Mountains over the granulite facies basement of the northeast Ox Mountains indicate that crustal thickening accompanied transcurrent deformation. It is however clear that the scale of crustal thickening produced by individual zones of transpression such as the Ox Mountains alone is small compared to that which produces large-scale granite genesis (cf Houseman et al. 1981). Figure 7.6 illustrates a typical cross–section showing typical scales of crustal thickening envisaged by these workers. They regard the entire lithosphere to be thickened by a factor of two and the zone of increased crustal thickening to extend for400km. Therefore, within a broad zone of crustal thickening the presence of relatively discrete zones of locally increased thickening due to transpression may promote the production of granitic melts by causing a critical stability limit to be exceeded. Thus shear zones can act as sites for the initiation of melting. It

is considered to be unlikely that individual structures such as the Ox Mountains Shear Zone alone could produce sufficient crustal thickening to suggest a causal link with granite genesis. This mechanism may however be feasible as part of the general crustal thickening associated with the Caledonian Orogeny.

7.3.3 Emplacement of granitic intrusions in middle–lower crustal level transcurrent shear–zones

One of the central problems concerning the emplacement of granitic bodies is the so–called space problem. Granitic emplacement in shear–zones has been the subject of much recent research, a considerable proportion of which has been specifically directed towards the space problem.

Hutton (1982) demonstrated that the Main Donegal Granite was emplaced at an early stage in the development of the shear zone, in an extensional cavity produced as the shear–zone split axially, and the north wall moved away. Thus emplacement took place in a transtensional enviroment, and was largely passive with the magma buoyancy force playing only a minor role in generating space. Guinberteau et al., (1987), described the syn–kinematic emplacement of the Mortagne Pluton into a pull–apart void, produced by an east–west orientated jog in the northwest–southeast trending shear–zone. Castro (1986), demonstrated that the Hercynian plutons in the Centra Extramadura batholith were emplaced in lensoid extension cracks orientated at 45° to the shearing direction. Natal'in et al. (1986), also described plutons in the Soviet Far East which occur as echelons, oblique and related to movement in the southern Yakutsh dextral fault system. Davis (1982), has described emplacement mechanisms for Pan African plutons related to the Nadj shear–zone of Saudi Arabia where passively emplaced cauldrons occur in the walls of the shear–zone and balloons and diapirs occur within the shear–zone itself.

Shaw (1980) examined the close relation between plutonism and intracontinental shear–zones. He emphasised the importance of dyke propagation in magma transport, and stated that extensional fractures that can act as ascent conduits for magma may be developed at depths of greater than 40km. He considered that continental crustal rocks can experience brittle behaviour when an interstitial fluid phase (which may be a high viscosity silicate melt) reduces the effective stress and that fracturing at deep levels due to regional stresses is the most effective mechanism for magma transport in ancient orogens. If at the same time anisotropy exists in the stress field induced by the regional deformation, extensional fractures can be produced readily. In this situation pure extensional fractures develop perpendicular to the minor effective stress and can propagate upwards into the crust and downwards into the mantle.

Hutton (1988), synthesised all of these data and concluded that because of anisotropies and viscosity contrasts within the continental crust, and the existence of pre–existing structures, shear systems in convergent, transcurrent and extensional systems will operate on a linked system of failure planes. Within this system, gaps and areas of dilation will open transiently or permanently at both upper and lower crustal levels. Thus in the operation of regional tectonics there need be no space problem.

While data from the above examples support this statement, the situation in the Ox Mountains is different, and suggests that permitted emplacement in purely extensional or dilational regimes may not be able to explain the emplacement of granitic magma in all shear–zones. The problem here is analogous to that of understanding the mechanism of granite emplacement within thrust systems, where again the tectonic environment is regionally strongly compressive in character

The present study of the geometry and kinematics of the metasedimentary envelope to the Ox Mountains Granodiorite and the contact between the metasediments and the pluton indicates that the ambient tectonic environment before the intrusion, and during and after the cooling of the intrusion was strongly transpressional in nature. In addition to this the sheets of different igneous lithology that occur within the pluton are orientated parallel to its margins.

Section 5.3.1 illustrates the pre–intrusion geometry of the metasedimentary envelope. This is dominated by by a steep tectonic fabric that is axial planar to upright F3 folds whose axes lie parallel to the regional stretching lineation and suggest transpressive deformation. The boundaries between lithological units are orientated parallel to this foliation. These data are consistent with the emplacement being controlled by these strongly developed planar anisotropies orientated parallel to the shear–zone boundary.

The absence of any evidence for ballooning strains within the Ox Mountains indicates that emplacement was not achieved forcefully. The present study of the geometry of the Ox Mountains (Chapters 4 & 5) suggests that one of two possibilities may account for emplacement within a regional tectonic environment which was predominantly transpressive in nature. While it is possible that the tectonic environment switched from transpressive pre–intrusion to transtensional or dilational shortly before and during intrusion, and then switched back to transpressional again, no evidence has been observed in the metasediments of the area to support this model. It is possible however that the granodiorite itself now occupies the site of this dilation and that all of the evidence for dilation within the metasediments has been destroyed. It has been demonstrated earlier that the structure of the Ox Mountains is dominated by a series of high strain zones which anastamose both laterally and vertically. Differential movement on these zones, with one moving at a different rate to others will produce displacement and hence strain gradients within the inlier. It is therefore possible that even at mid–crustal depths, areas of relatively low transpressive strain or even local areas of dilational strain may exist. The present author considers that the granodiorite is likely to have been emplaced by intrusion of a series of sheets into these areas within a shear zone which was predominantly transpressive in character. The emplacement mechanism suggested by these data in the Ox Mountains may have relevance to the emplacement of granites within thrust systems where again the overall tectonic environment is strongly compressive but where displacent gradients and therefore areas of locally dilational strain are likely to occur.

7.4 CONCLUSIONS

1. The geometry of shear zones at deep levels is poorly understood. Data from the Ox Mountains suggest that at middle–lower crustal levels they may be expressed as an anastamosing series of high strain zones that separate areas of lower tectonic strain. The geometry of these high strain zones indicates that the shear zone broadens with depth. This geometry and the presence of a shallowly inclined basement–cover interface in the north of the area is consistent with this representing a shallowly inclined décollement surface. These data do not preclude a strand or strands of the braided fault system extending deeper into the crust or even mantle to form a conduit along which magma may be introduced into the system.

2. The presence of a décollement at the basement–cover interface within a strike–slip fault system introduces the concept of basement stripping, which is a viable mechanism for the shuffling and rotation of terranes and the removal of slivers of basement material along the axis of the fault system.

3. It is considered unlikely that individual shear zones such as the Ox Mountains Shear Zone could produce crustal thickening on a sufficiently large scale to suggest a causal link with granite genesis. Structures such as the Ox Mountains Shear Zone will however contribute to the overall crustal thickening present within convergent zones and may therefore act as favourable sites for the initiation of melting.

4. The close spatial relationship observed between granitic intrusions and shear zones is therefore considered to be due largely to shear zones acting as favourable conduits for the ascent of magma.

5. Data from the Ox Mountains suggest that granite emplacement may be achieved by magma exploiting localized areas of low transpressive or even dilational strain produced by displacement gradients within a tectonic environment which is overall transpressive in character. A similar emplacement mechanism may account for the emplacement of granitic intrusions into thrust systems.

CHAPTER 8

CONCLUSIONS

8.1 CHAPTER 2. The Ox Mountains tectono–stratigraphic sequence

Lithostratigraphic and structural data from the Ox Mountains have been used to develop a tectono-stratigraphic sequence. Way–up criteria observed in the Leckee Quartzitic Formation and the Cloonygowan Formation cannot be extrapolated into the rest of the sequence due to the presence of tectonic slides. Identification of the Port Askaig Tillite within the Ox Mountains Succession is of critical importance as it provides a unique lithostratigraphic marker horizon, confirms that the metasediments of the central and southwest Ox Mountains are of Dalradian age, and enables correlation to be made with other areas of Dalradian rocks in which the stratigraphic succession is more clearly defined.

Way–up criteria within the Leckee Quartzitic Formation indicates that this unit lies stratigraphically above the Port Askaig Tillite. Identification of the Leckee Transition Member, which represents the transition between the Quartzite and the Ummoon Formation, establishes an important stratigraphic link between these two formations, which are normally separated by the Lough Talt Slide. This also establishes that the Ummoon Formation is a direct lateral equivalent of the pelite, semi-pelite and psammite exposed on the northwestern side of the granodiorite (Attymass Group after Taylor, (1968); Ellaghmore, Killgellia and Carrick O' Hara Formations after Andrews et al. (1978)), which structurally overlies the Leckee Transition Member.

The criteria cited by Winchester et al. (1988), to distinguish the Ummoon Formation from the Lismoran Formation are not accepted. The present author has retained the arbitrary division between the two formations proposed by Phillips et al. (1975), and considers that the Ummoon Formation passes into the Lismoran Formation via a sedimentary transition. There is therefore no stratigraphic basis for regarding the Glennawoo Slide as a major terrane boundary.

The discovery of large elliptically shaped volcanic clasts within the Callow Member, following the identification of volcanic bombs within the Newantrim Member (Long and Max, 1977), confirms that both units are of volcanic origin.

The stratigraphic relationship of the greenschist facies rocks of the Cloonygowan formation to the higher grade rocks of the Ox Mountains Succession is unknown. At Cloongowan and Ardvarney an exposure gap of a minimum of 25m separates the two units which are lithologically distinct. The conclusion of Phillips et al. (1975), that the boundary between them is a gradational and probably interbanded one is therefore not accepted by the present author.

The difference in the lithostratigraphic and metamorphic history of the Cloonygowan Formation and the Ox Mountains Succession is consistent with the Cloonygowan Formation representing a separate allochthonous terrane. Structural data however indicates that both units have experienced an identical deformation history, which suggests they were in close proximity during their deformation. This can be explained if the Cloonygowan Formation was deformed at a higher level in the strike slip system and was then lowered into its present position along a late fault, following the peak of regional metamorphism.

8.2 CHAPTER 3. Correlation with the Dalradian Succession of southern Donegal; the Ox Mountains stratigraphic succession.

The close correlation between the Ox Mountains metasediments and those of the Dalradian succession in S. Donegal, whose way–up and age are known, strongly suggests that the tectono–stratigraphic sequence established in the Ox

Mountains represents the stratigraphic succession. The metasediments of the Ox Mountains sequence are therefore referred to as the Ox Mountains Succession.

The correlation with the upper part of the southern Donegal Succession, which is stratigraphically continuous, ie not dismembered by major tectonic slides, is inconsistent with the model of terrane accretion proposed by Winchester, Long and Max (1988), which necessitates a major terrane boundary to separate the Ummoon and Lismoran formations in the Ox Mountains.

8.3 CHAPTER 4. Deformation chronology

The rocks of the Ox Mountains Succession are considered to have experienced a similar structural history to those of the Cloonygowan Formation although at a deeper structural level.

S1 is rarely preserved and has been positively identified only in the hinges of F2 folds, which are themselves uncommon. No evidence has been observed to suggest that any major structures related to D1 are present.

D2 was an important period in the structural history of the inlier. A flat lying penetrative fabric (S2) was developed and occasional F2 folds formed. The peak of regional metamorphism at the kyanite zone of the amphibolite facies was attained following D2 (MP2). Transposition by D3 structures conceals much of the original geometry of D2.

The present geometry of the metasediments of the central and southwest Ox Mountains largely results from the progressive sinistral transcurrent deformation. D3 produced a moderate to steeply inclined northeast–southwest trending fabric (S3), axial planar to gently plunging F3 mesoscopic folds which verge towards a major F3 antiform that forms the core of the inlier. The Ox Mountains Granodiorite was synkinematically intruded into the core of this structure, close to the amphibolite facies peak of regional metamorphism. Continuing transcurrent deformation affected the inlier and produced sinistral shear fabrics both within the pluton and in the metasedimentary envelope where the S3 fabric locally intensifies towards tectonic slides of D3 age. The gently south–dipping contact between the Dalradian metasediments of the central and southwest Ox Mountains and the granulite facies psammites of the northeast Ox Mountains has also been identified as a tectonic slide of D3 age. The geometry of the Ox Mountains is therefore dominated by a series of D3 high strain zones which converge both laterally and vertically. As the metamorphic grade decreased the major slides locked up. The Knockaskibole fault then became active as a break–back structure.

Transcurrent deformation ceased following D3 and uplift began throughout the inlier. The geometry and timing of the development of D4 structures suggests that they could, at least in part, represent this fundamental change in kinematics. Metamorphic grade at this stage did not exceed the greenschist facies.

Renewed sinistral deformation (D5) occurred synchronously with or slightly later than the emplacement of the Lough Easky and Lough Talt Adamellites.

8.4 CHAPTER 5. A tectono–thermal model for the Ox Mountains

The structural and kinematic analysis of chapter 4 illustrates that the inlier has experienced a complex structural history. Four main kinematic episodes are represented.

1. Initial fold and fabric development, and activity on the contact between the Cloonygowan Formation and the Ox Mountains Succession, (D1–D2, Pre-478 Ma).

2. Sinistral transcurrent deformation, development of a braided fault system and syn-kinematic intrusion of the Ox Mountains granodiorite under amphibolite fa-

cies conditions at 478±12Ma (D3, Arenig-Llanvirn), followed by further sinistral deformation as the metamorphic grade decreased and the granodiorite cooled.

3. Uplift, reflected by decreasing metamorphic grade (D4).

4. Sinistral transcurrent deformation following intrusion of the Lough Talt Adamellite and Lough Easky Adamellite under greenschist facies conditions at 401±33Ma (D5, Early Devonian).

8.5 CHAPTER 6. Regional tectonic synthesis and correlation

The protracted history of sinistral deformation recorded by the Ox Mountains can be related to terrane accretion events along the Highland Boundary Fault Zone. This provides information on the early kinematic history of this structure that is not available in the remainder of the British Isles.

During the early Ordovician the Ox Mountains and the Clew Bay region both experienced intense sinistral deformation. The timing of deformation in the former area is constrained by the synkinematic Ox Mountains Granodiorite, dated at 478±12 Ma. In the latter area the age of deformation is constrained to be pre–Silurian, as Silurian sediments lie unconformably upon sinistrally–sheared serpentinites.

This initial period of sinistral deformation is succeeded by a period of uplift, whose timing is constrained to be between the synkinematic emplacement of the Ox Mountains Granodiorite and the emplacement of the adamellites at a higher crustal level. This uplift in the Clew Bay – Ox Mountains region is correlated with the widely recognised uplift that occurred throughout the Dalradian between 460 and 440 Ma.

The Silurian was a period of relative tectonic inactivity. Sinistral deformation is recorded in the Silurian rocks of the South Mayo Trough, where the Croagh

Patrick syncline developed due to sinistral transpression during post–Wenlock to Siegennian times. The Clew Bay Thrust was reactivated as a sinistral structure at this time. Sinistral deformation is also recorded by the deformation of adamellites in the Ox Mountains dated at 401±12 Ma. This second trancurrent event is comparatively minor relative to the early Ordovician period of intense sinistral strike–slip deformation. Data from the remainder of the British Isles suggest that the majority of the strike slip deformation associated with terrane accretion occurred at about 400 Ma (Hutton, 1987). The absence of evidence for intense strike slip deformation within the Ox Mountains at this time is consistent with the hypothesis that the main locus of strike slip deformation had shifted elsewhere, possibly to the southeast.

The deposition of relatively undeformed Devonian sediments unconformably upon the Silurian places an upper limit on the age of deformation in this sector of the Highland Boundary Fault Zone.

8.6 CHAPTER 7. The geometry and kinematics of the Highland Boundary Fault in western Ireland: implications for the structural geometries of accretionary tectonics

The geometry of shear–zones at deep levels is poorly understood. Data from the Ox Mountains suggests that at middle–lower crustal levels they may be expressed as an anastamosing series of high strain zones that separate areas of lower tectonic strain. The geometry of these high strain zones indicates that the shear zone broadens with depth. This geometry and the presence of a shallowly inclined basement–cover interface in the north of the area is consistent with this representing a shallowly inclined decollement surface. These data do not preclude a strand or strands of the braided fault system extending deeper into the crust or even mantle to form a conduit along which magma may be introduced into the system.

The presence of a décollement at the basement–cover interface within a strike–slip fault system introduces the concept of basement stripping, which is a viable mechanism for the shuffling and rotation of terranes and the removal of slivers of basement material along the axis of the fault system.

It is considered unlikely that individual shear zones such as the Ox Mountains Shear Zone could produce crustal thickening on a sufficiently large scale to suggest a causal link with granite genesis. Structures such as the Ox Mountains Shear Zone will however contribute to the overall crustal thickening present within convergent zones and may therefore act as favourable sites for the initiatation of melting.

The close spatial relationship observed between granitic intrusions and shear zones is therefore largely considered to be due to shear zones acting as favourable conduits for the ascent of magma.

Data from the Ox Mountains suggests that granite emplacement may be achieved by magma exploiting localized areas of low transpressive or even dilational strain produced by displacement gradients within a tectonic environment which is overall transpressive in character. A similar emplacement mechanism may account for the emplacement of granitic intrusions into thrust systems.

REFERENCES

ALLEN, C. R. 1957. San Andreas fault zone in San Gorgonio Pass, Southern California. Bulletin of the Geological Society of America **68**, 315-350.

ALLEN, C. R. 1965. Transcurrent faults in continental areas. In: A symposium on continental drift: Philosophical transactions of the Royal Society of London, **1088**, 82-89.

ALSOP, G. I. 1987. The Dalradian of Central Donegal, Ireland; an example of a polyphase mid-crustal thrust zone. Unpublished Ph.D Thesis (University of Durham).

ANDERTON, R. I. 1985. Sedimentation and tectonics in the Scottish Dalradian. Scottish Journal of Geology, **21**, 407-436.

ANDREWS, J. R. 1984. Fracture controlled feldspar shape fabrics in deformed quartzo-feldspathic rocks. Journal of Structural Geology, **2**, 175-187.

ANDREWS, J. R., PHILLIPS, W. E. A. & MOLLOY, M. A. 1978. The metamorphic conditions of part of the north central Ox Mountains inlier of Counties Sligo and Mayo. Journal of Earth Science Royal Dublin Society, **1**, 173-194.

ARMSTRONG, R. L. & DICK, H. J. B. 1974. A model for the development of thin overthrust sheets of crystalline rock. Geology, **2**, 35-40.

BAILEY, E.B. & HOLTEDAHL, O. 1938. Northwest Europe Caledonides, In: Regionale geologie der Erde. Part 2: Leipzig Akademie. Verlag, 84pp.

BAK, J., GROCOTT, J., KORSTGARD, J.A., SORENSON, K., NASH, D. F. & WATTERSON, J. 1975. Tectonic implications of Precambrian shear belts in western Greenland. Nature, London, **254**, 566–9.

BALLY, A. W., 1981. Thoughts on the tectonics of folded belts. In: McClay, K. A. A & Price, N. J. (editors) Thrust and Nappe Tectonics. Special Publication of the Geological Society, London, **9**, 13–32.

BALLY, A. W. 1983. Seismic expression of structural styles: AAPG studies in Geology, series 15, 3 volumes.

BARTLETT, W. L., FRIEDMAN, M., & LOGAN, J. M. 1981. Experimental folding and faulting of rocks under confining pressure, Part 9. Wrench faults in limestone layers: Tectonophysics **79**, 255–277.

BEACH, A. 1980. Retrogressive metamorphic processes in shear zones. Journal of Structural Geology, **2**, 257-263.

BELL, T. H. & ETHERIDGE, M. A. 1973. Microstructures of mylonites and their descriptive terminology. Lithos, **6**, 337-348.

BERBERIAN, M. 1981. Active faulting and tectonics of Iran. In Gupta, H. K. and Delany, F. M. editors. Zagros, Hindu Kush, Himalaya Geodynamic evolution: Geodynamics series 3: Washington D.C., and Boulder Colorado. American Geophysical Union and the Geological Society of America, 33–49.

BERTHE, D., CHOUKROUNE, P. & GAPAIS, D. 1979a. Orientations preferentielles du quartz et orthogneissification progressive en regime cisaillement: l'example du cisaillement Sudarmoricain. Bulletin de Mineralogie, **102**, 265-272.

BERTHE, D., CHOUKROUNE, P. & JEGOUZO, P. 1979b. Orthogneiss mylonite and non-coaxial deformation of granites: The example of the South Armorican Shear Zone. Journal of Structural Geology, **1**, 31-42.

BICKLE, M. J., KIDD, R. G. W. & NISBET, E. 1972. The Silurian of the Croagh Patrick Range, County Mayo. Scientific proceedings, The Royal Society, Dublin, Series A, 231–249.

BLISS, G. M. 1986. The Micropalaeontology of the Dalradian. Unpublished PhD thesis, The University of London.

BLUCK, B. J. 1984. Pre–Carboniferous history of the Midland Valley of Scotland. Transactions of the Royal Society, Edinburgh, Earth Sciences, **75**, 275–295.

BLUCK, B. J. 1985. The Scottish paratectonic Caledonides. Scottish Journal of Geology, **21**, 437–464.

BORRADAILE, G. J. 1973. Dalradian structure and stratigraphy of the Lough Aire district, Argyllshire. Transactions of the Royal Society of Edinburgh, **69**, 1-21.

BORRADAILE, G. J. 1974. The structures of the Moine-like rocks near Lough Derg, County Donegal, Eire. Geological Journal, **9**, 61-64.

BREWER, J. A. & SMYTHE, D. K. 1984. MOIST and the continuity of reflector geometry along the Caledonian–Appalachian orogen. Journal of the Geological Society, London, **41**, 105–120.

BRIDWELL, R. J. 1975. Sinuosity of strike–slip fault traces: Geology, **3**, 630–632.

BRUN, J. P. & PONS, J. 1981. Strain patterns and pluton emplacement in a crust undergoing non–coaxial deformation. Journal of Structural Geology, **3**, 219–229.

BURKE, D. B. 1979. Log of a trench in the Garlock Fault zone, Fremont Valley, California: USGS Map MF–1028, scale 1:20.

BUTLER, R. W. H. 1987. Thrust sequences. Journal of the Geological Society, London, **144**, 6199–634.

CASTRO, A. 1985. The Central Extremadura Batholith: Geotectonic implications (European Hercynian Belt) - an outline. Tectonophysics, **120**, 57-68.

CASTRO, A. 1987. On granitoid emplacement and related structures: a review. Geologische Rundschau **76/1**, 101–124.

CHEADLE, M. J., et al. 1986. The deep structure of the Mojave desert California, from COCORP seismic reflection data: Tectonics, **5**, 293–320.

CHEN, W. P., & MOLNAR, P. 1983. Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere. Journal of Geophysical Research, **88**, 4183–4214.

CLAYTON, L. 1966. Tectonic depressions along the Hope fault, a transcurrent fault in north Canterbury, New Zealand. New Zealand Journal of Geology and Geophysics, **9**, 95–104.

COGNE, J., DURAND, S. & ESTOUTE-CHOUX, J. 1974. Le Massif Amoricain. In: Debelmass, J. (editor) Geologie de la France, Dois, Paris, **1** , 105–161.

COLE, G. A. J., KILROE, J. R., HALISSY, T. & NEWELL ARBER, E. A. 1914. The geology of Clare Island, County mayo. Memoir of the Geological Survey of Ireland.

CONEY, P. J. 1989. Structural aspects of suspect terranes and accretionary tectonics in western North America. Journal of Structural Geology, **11**, 107–125.

CONEY, P. J. & JONES, D. L. 1985. Accretion tectonics and crustal structure in Alaska. Tectonophysics, **119**, 265–283.

CRADDOCK, C., HAUSER, E. C., MAHER, H. D., SUN, A. Y., & ZHU, G. Q. 1985. Tectonic evolution of the West Spitzbergen fold belt. Tectonophysics, **114**, 193–211.

CURRALL, A. E. 1963. The geology of the southwest end of the Ox Mountains, County Mayo. Proceedings of the Royal Irish Academy, **63B**, 131.

CURRALL, A. E. & TAYLOR, W. E. G. 1965. Geological history of the Slieve Gamph mountain, Western Ireland. Nature, **207**, 1143-1144.

CURRY, G. B., INGHAM, J.K., BLUCK, B. J. & WILLIAMS, A. 1982. The significance of a reliable Ordovician age for some Highland Border rocks in central Scotland. Journal of the Geological Society, London, **139**, 453-456.

CURRY, G. B., BLUCK, B. J., BURTON, C. J. INGHAM. J. K., SIVETER, D. J. & WILLIAMS, A. 1984. Age, evolution and tectonic history of the Highland Border Complex, Scotland. Transactions of the Royal Society, Edinburgh : Earth Sciences, **75**, 113–133.

DAVIS, T. L. & DUEBENDORFER, E. M. 1987. Strip map of the western big bend segment of the San Andreas fault: Geological Society of America Map and Chart Series, Map 60, Scale 1:31,682.

DEMPSTER, T. J. 1983. Studies of orogenic evolution in the Scottish Dalradian. Unpublished Ph.D. thesis, Edinburgh University.

DEMPSTER, T. J. 1984a. Localised uplift in the Scottish Dalradian. Nature, **307**, 156–159.

DEMPSTER, T. J. 1985. Uplift patterns and orogenic evolution in the Scottish Dalradian. Journal of the Geological Society, London, **142**, 111–128.

DEWEY, J. F. 1962. The provenance and emplacement of Upper Arenigian turbidites in Co. Mayo, Eire. Geological Magazine, **99**, 238–252.

DEWEY, J. F. 1982. Plate tectonics and the evolution of the British Isles. Journal of the Geological Society, London, **139**, 371–412.

DEWEY, J. F. & PHILLIPS, W. E. A. 1963. A tectonic profile across the Caledonides of South Mayo. Liverpool and Manchester Geological Journal, **3**, 237-246.

DEWEY, J. F. & PANKHURST, R. J. 1970. The evolution of the Scottish Caledonides in relation to their isotopic age pattern. Transactions of the Royal Society of Edinburgh, **68**, 362–389.

DEWEY, J. F., McKERROW, W. S. & MOORBATH, M. 1970. The relationship between isotopic ages, uplift and sedimentation during Ordovician times in western Ireland. Scottish Journal of Geology, **6**, 133–145.

DEWEY, J. F. & SHACKLETON, R. M. 1984. A model for the evolution of the Grampian tract in the early Caledonides and Appalachians. Nature, **312**, 115-121.

DEWEY, J. F., HEMPTON, M. R., KIDD, W. S. F., SAROGLU, F. & SENGOR, A. M. C. 1986. Shortening of the continental lithosphere: The neotectonics of eastern Anatolia – a young collision zone, in: Coward, M. P. & Ries, A. C., editors, Collision tectonics: Special Publication of the Geological Society of London, **19**, 3–36.

DICKINSON, W. R. 1983. Cretaceous sinistral strike–slip along Nacimento fault in central California. Bulletin of the American Association of Petroleum Geologists, **67**, 624–645.

DUNNING, F. W., TROTTER, F. W. & BOGDANOFF, A. 1965. International tectonic map of Europe: Sheet 5, Scale 1:2,500,000.

DUNNING, F. W. & MAX, M. D. 1975. Tectonic map of Ireland.

ELWELL, R. W. D. 1955. The lithology and structure of a boulder-bed in the Dalradian of Mayo, Ireland. Quarterly Journal of the Geological society of London, **111**, 71-84.

FITTON, J. G. & DUNLOP, H. M. 1985. The Cameroon Line, West Africa, and its bearing on the origin of oceanic and continental alkali basalts. Earth and Planetary Science Letters, **72**, 23–38.

FLINN, D. 1962. On folding during three dimensional progressive deformation. Journal of the Geological Society, London, **118**, 385–423.

FLEUTY, M. J. 1964. The description of folds. Proceedings of the Geological Association, **75**, 461-489.

FREUND, R. 1974. Kinematics of transform and transcurrent faults. Tectonophysics, **21**, 93–134.

GAPAIS, D. & LE CORRE, C. 1980. Is the Hercynian belt of Brittany a major shear zone ? Nature, London, **88**, 547–566.

GIEKIE, A. 1893. Annual Report of the Geological Survey and Museum of Practical Geology, Appendix E.

GIBBONS, W. & GAYER, R. A. 1985. British Caledonian Terranes. In: Gayer, R. A. (editor). The tectonic evolution of the Caledonian–Appalachian Orogen, Viewseg, Brunswick, 3–16.

GRAHAM, J. R. 1981. The ORS of Co. Mayo, Northwest Ireland. Geological Journal, **16**, 157–173.

GRAHAM, J.R. 1983. The Devonian of the Ox Mountains. In: J. B. Archer & P. D. Ryan (Editors.). Geological guide to the Caledonides of Western Ireland. Guide Series of the Geological Survey of Ireland, **4**, 55-58.

GRAHAM, J. R. & SMITH, D. G. 1981. The age and significance of a small Lower Palaeozoic inlier in County Mayo. Journal of Earth Sciences, The Dublin royal society, **4**, 1–5.

GRAHAM, J. R., LEAKE, B. E. & RYAN, P. D. 1985. The Geology of South Mayo. 1:63,360 Map. University of Glasgow.

GRAHAM, J. R. RICHARDSON, J. B. & CLAYTON, G. 1983. Age and significance of the ORS around Clew Bay, NW Ireland. Transactions of the Royal Society of Edinburgh: Earth Sciences, **73**, 245–249.

GROSE, L. T. 1959. Structure and petrography of the northeast part of the Soda Mountains, San Bernardino County, California. Bulletin of the Geological Society of America, **70**, 1509–1548.

GUINEBERTEAU, B., BOUCHEZ, J. L. & VIGNERESSE, J. L. 1987. The Mortagne granite pluton (France) emplaced by pull-apart along a shear zone: structural and gravimetric arguments, regional implication. Bulletin of the Geological Society of America, **99**, 763-770.

HADLEY, D. & KANAMORI, H. 1977. Seismic structure of the Transverse Ranges, California: Bulletin of the Geological Society of America, **88**, 1469–1478.

HARDING, T. P. & LOWELL, J. D. 1979. Structural styles, their Plate–Tectonic habitats and hydrocarbon traps in petroleum provinces. Bulletin of the American Association of Petroleum Geologists, 63, 1016–1058.

HARDING, T. P. 1985. Seismic characterisation and identification of negative flower structures, positive flower structures and positive structural inversion. Bulletin of the American Association of Petroleum Geologists, **69**, 582–600.

HARLAND, W. B. 1971. Tectonic transpression in Caledonian Spitsbergen. Geological Magazine, **108**, 27–42.

HARPER, D. A. T., WILLIAMS, D. M. & ARMSTRONG, H. A. 1989. stratigraphic correlations adjacent to the Highland Boundary Fault in the west of Ireland. Journal of the Geological Society, London, **146**, 381–384.

HARRIS, A. L. & PITCHER, W. S. 1975. The Dalradian Supergroup. Special Report of the Geological Society, London, **6**, 52-75.

HARRIS, A. L., BRADBURY, H. J. & McGONIGAL, M. H. 1976. The evolution and transport of the Tay Nappe. Scottish Journal of Geology, **12**, 103–113.

HARTE, B., & JOHNSTONE, M. R. W. 1969. Metamorphic history of the Dalradian rocks in Glens Clova, Esk and Lethnot, Angus, Scotland. Scottish Journal of Geology, **5**, 54–80.

HARTE, B., DEMPSTER, T. J. & BOOTH, J. E. 1984. P–T estimates and mineral facies variation in space and time in the central and southeast Scottish Highlands. Journal of the Geological Society, London, **142**, 58–71.

HARTE, B., BOOTH, J. E., DEMPSTER, T. J., FETTES, D. J., MENDUM, J. R. & WATTS, D. 1984. Aspects of the post–depositional evolution of Dalradian and Highland Border Complex rocks in the Southern Highlands of Scotland. Transactions of the Edinburgh Royal Society, **75**, 466–477.

HEARNE, T. M. & CLAYTON, R. W. 1986A. Lateral velocity variations in Southern California. 1, results for the upper crust from Pg waves. Bulletin of the Seismological Society of America, **76**, 495–509.

HEARNE, T. M. & CLAYTON, R. W. 1986B. Lateral velocity variations in southern California. II, results for the upper crust from Pn waves. Bulletin of the Seismological Society of America, 76, 511–520.

HOUSEMAN, G. A. McKENZIE, D. P. & MOLNAR. P. 1981. Convective instablity of a thickened boundary layer and its relevance to the thermal evolution of continental convergent belts. Journal of Geophysical Research, **86**, B7, 6115–6132.

HOUSEMAN, G. A. & ENGLAND, P. C. 1986. A dynamical model of lithosphere extension and sedimentary basin formation. Journal of Geophysical Research, **91**, 719–729.

HOWARTH, R.J. 1971. The PortasKaig Tillite Succession (Dalradian) of County Donegal. Proceedings of the Royal Irish Academy, **71B**, 1-36.

HOWARTH, R. J., KILBURN, C. & LEAKE, B. E. 1966. The Boulder-Bed Succession at Glencolumbhille, County Donegal. Proceedings of the Royal Irish Academy, **65B**, 156.

HOWELL, D. G., JONES, D. L. & SCHERMER, E. R. 1985. Tectonostratigraphic Terranes of the Circum–Pacific Region, In: Tectonostratigraphic terranes of the Circum–Pacific Region. The Circum–Pacific Council for Energy and Mineral Resources.

HUTTON, D. H. W. 1979b. Tectonic Slides: a review and reappraisal. Earth Science Reviews, **15**, 151–172.

HUTTON, D. H. W. 1982a. A tectonic model for the emplacement of the Main Donegal granite, northwest Ireland. Journal of the Geological Society, London, **139**, 615-631.

HUTTON, D. H. W. 1987. Strike–slip terrane accretion model for the evolution of the British and Irish Caledonides. Geological Magazine, **124**, 405–425.

HUTTON, D. H. W. 1988. Granite emplacement mechanisms and tectonic controls. Transactions of the Royal Society of Edinburgh, **79**, 245-255.

HUTTON. D. H. W. & DEWEY, J. F. 1986. Palaeozoic terrane accretion in the western Irish Caledonides. Tectonics, **5**, 1115–1124.

JACKSON, J. A., GAGNEPAIN, J., HOUSEMAN, G., KING, G., PAPADIMITRIOU, P., SOUFLERIS, C. & VIRIEUX, J. 1982. Seismicity, normal faulting and the geomorphical evolution of the Gulf of Corinth: the Corinth earthquakes of February and March 1981. Earth and Planetary Science Letters, **57**, 377–397.

JACKSON, J. A. 1987. Active normal faulting and crustal extension. In: Coward, M. P., Dewey, J. F. & Hancock, P. L. (editors), Continental Extensional Tectonics, Special publication of the Journal of the Geological Society, London, **28**, 3–17.

JEGUSO, P. 1980. The South Armorican Shear zone. Journal of Structural Geology, **2**, 39–47.

JOHNSTONE, G. S. 1975. The Moine Succession. In: A correlation of Precambrian rocks in the British Isles. Special Report of the Geological Society of London, **6**, 30-42.

JOHNSON, M.R. W. & HARRIS, A. L. 1967. Dalradian–? Arenig relations in parts of the Highland Border, Scotland, and their significance in the Caledonian orogeny. Scottish Journal of Geology, **3**, 1–16.

JOHNSTONE, G. S., SMITH, D. I. & HARRIS, A. L. 1969. The Moinian Assemblage of Scotland. In: Kay, M. (Ed.). North Atlantic - geology and continental drift, a symposium. Memoir of the American Association of Petroleum Geologists, **12**, 159-180.

JONES, C. S. & LEAT, P. T. 1988. Discussion on trace element geochemical correlation in the reworked Proterozoic Dalradian metavolcanic suites of the western Ox Mountains and Northwest Mayo Inliers, Ireland. Journal of the Geological Society, London, **145**, 1037–1040.

JONES, D. L., HOWELL, D. G., CONEY, P. J. & MONGER, J. W. H. 1983. Recognition, character and analysis of tectonostratigraphic terranes in western North America. In: Hasimoto, M. & Ugeda, S. (editors) Accretion tectonics in the Circum–Pacific Regions, Terra publications, Tokoyo, 21–35.

KELLOGG, H. E. 1975. Tertiary stratigraphy and tectonism in Svalbard and continental drift. Bulletin of the American Association of Petroleum Geologists, **59**, 465–485.

KENNEDY, M. J. 1969. The structure and stratigraphy of the Dalradian rocks of north Achill Island, County Mayo, Ireland. Quarterly Journal of the Geological Society of London, **125**, 47-81.

KILBURN, C., PITCHER, W.S. & SHACKLETON, R. M. 1965. The structure and origin of the Portaskaig Boulder-Bed Succession (Dalradian). Geological Journal, **4**, 343-360.

KINAHAN, G. H. 1882. Palaeozoic rocks of Galway and elsewhere in Ireland, said to be Laurentians. Scientific Proceedings of the Royal Society, Dublin, **3**, 347–356.

KINGMA, J. T. 1958. Possible origin of piercement structures, local unconformities and secondary basins in the Eastern Geosyncline, New Zealand. New Zealand Journal of Geology and Geophysics, **1**, 269–274.

KNIPE, R. J., 1989. Deformation mechanisms – recognition from natural tectonites. Journal of Structural Geology, **11**, 127–146.

LACHENBRUCH, A. H. & SASS, J. H. 1977. Heat flow in the United States and the thermal regime in the crust. Monograph of the American Geophysical Union, **20**, 626–675.

LAPWORTH, C. 1885. The Highland controversy in British geology: its causes and consequences. Nature, **32**, 558-559.

LEMINSZKI, P. J. & BROWN, L. D. 1988. Variable crustal structure of strike slip fault zones as observed on deep seismic reflection profiles. Geological Society of America, Bulletin, **100**, 665–676.

LEMON, G. G. 1966. Serpentinisation in the northeast Ox Mountains, Eire. Geological Magazine, **103**, 124-137.

LEMON, G. G. 1971. The Precambrian rocks of the northeast Ox Mountains, Eire. Geological Magazine, **89**, 124-128.

LISTER, G. S. & SNOKE, A. W. 1984. S-C Mylonites. Journal of Structural Geology, **6**, 617-638.

LONG, C. B. 1974. A note on the stratigraphy of the probable Dalradian metasediments north and northeast of Castlebar, County Mayo. Bulletin of the Geological Survey of Ireland, **1**, 459-469.

LONG, C. B. 1983. The Lough Anaffrin Area (Central Ox Mountains Inlier). In: J. B. Archer & P. D. Ryan (Editors). Geological Guide to the Caledonides of Western Ireland. Guide Series of the Geological Survey of Ireland, **4**, 47-50.

LONG, C. B., FLEGG, A. M. & CRILLY, K. 1981. Lough Anaffrin Marble Prospect, near Castlebar, County Mayo. Open File report of the Geological Survey of Ireland.

LONG, C. B. & MAX, M. D. 1977. Metamorphic rocks in the southwest Ox Mountains Inlier, Ireland; their structural compartmentation and place in the Caledonian orogen. Journal of the Geological Society of London, **133**, 413-432.

LONG, C. B., MAX, M. D. & O'CONNOR, P. J. 1984. Age of the Lough Talt and Easby adamellites in the central Ox Mountains, northwest Ireland, and their structural significance. Geological Journal, **19**, 389-397.

LONG, C. B. & YARDLEY, B. W. D. 1979. The distribution of pre-Caledonian Basement in the Ox Mountains Inlier, Ireland. In: A. L. Harris et al (Editors). The Caledonides of the British Isles - reviewed. Geological Society of London Special Publication, **8**, 153-156.

LONG, C. B., MAX, M. D. & YARDLEY, B. W. D. 1982. Compilation metamorphic map of Ireland. In: N. A. T. O. Advanced Study Institute on Regional trends, A. M. H. 425, P. E. Schenk (Editor).

LOWELL, J. D. 1969. On the origin of upthrusts. Geological Society of America, Abstracts with Programs for 1969, Part 7 (annual meeting), 138.

LOWELL, J. D. 1972 Spitzbergen Tertiary Orogenic belt and the Spitzbergen Fracture Zone. Geological Society of America, Bulletin, **83**, 3091–3102.

LUYENDYK, B. P., KAMERLING, M. J. TERRES, R. R., & HORNAFIUS, J. S. 1985. Simple shear of southern California during Neogene time suggested by palaeomagnetic declinations. Journal of Geophysical Research, **90**, (B14), 12454–12466.

MAX, M. D. 1975. In: F. W. Dunning & M. D. Max. Explanatory notes to the geological map of the exposed and concealed Precambrian basement of the British Isles. In: A correlation of Precambrian rocks in the British Isles. Special Report of the Geological Society of London, **6**, 11-14.

MAX, M. D. 1982. Deformation in the Irish Caledonides. In: N. A. T. O. Advanced Study Institute on Regional trends, A. M. H. **425**, P. E. Schenk (Editor).

MAX, M. D. 1989. The Clew Bay Group: A Displaced Terrane of the Highland Border Group Rocks (Cambro–Ordovician) in Northwest Ireland. Geological Journal, **24**, 1–19.

MAX, M. D. & INAMDAR, D. D. 1983. Detailed compilation magnetic map of Ireland and a survey of its deep geology. Report R. S. 82/2, Geological Survey of Ireland, Dublin.

MAX, M. D. & LONG, C. B. 1977. Discussion of the geological effects of the Slieve Gamph Igneous Complex, County Mayo, Ireland. Journal of the Geological Society of London, **133**, 395.

MAX, M. D. & LONG, C. B. 1979. Basic volcanic rocks in the Dalradian of Ireland. In: Harris, A. L., Holland, C. H. & Leake. B. E. (editors) The Caledonides of the British Isles Reviewed. Special Publication of the Geological Society, London, number 8, 585-591.

MAX, M. D. & LONG, C. B. 1985. Pre–Caledonian basement in Ireland and its cover relationships. Geological Journal, **20**, 341–346.

MAX, M. D., LONG, C. B. & SONET, J. 1976. The geological setting and age of the Ox Mountains Granodiorite. Bulletin of the Geological Survey of Ireland, **2**, 27–35.

MAX, M. D., O"CONNOR, P. J. & LONG, C. B. 1984. New age data for the Pre–Caledonian basement of the northeast Ox Mountains and Lough Derg Inliers, Ireland. Bulletin of the Geological Survey of Ireland, **3**, 203–209.

MAX, M. D. & RIDDIHOUGH, R. P. 1975. Continuation of the Highland Boundary Fault in Ireland. Geology, **3**, 206–210.

MAX, M. D., RYAN, P. D. & INAMDAR, D.D. 1983. A magnetic deep structural geology interpretation of Ireland. Tectonics, **2**, 431–451.

Mc HENRY, A. 1903. Report on the Ox Mountains rocks and their probable continuation from Galway and Mayo into Donegal, Tyrone and Londonderry. Proceedings of the Royal Irish Academy, **24**, 371–378.

MEISSNER, R. & WEVER, T. 1986. Intracontinental seismicity, strength of crustal units and the seismic signature of fault zones. Philosophical Transactions of the Royal Society of London. **A 317**, 45–61.

MOLLOY, M. A. & SANDERS, I. S. 1983. The Coolaney area. In: Archer, J. B. & Ryan. P. D. (editors) Geological Guide to the Caledonides of Western Ireland. Geological Survey of Ireland Guide Series, **4**, 52–55.

MOORE, J. McM. 1979. Tectonics of the Najd transcurrent fault system, Saudi Arabia. Journal of the Geological Society, London, **136**, 441–454.

MORTON, A. C. & TAYLOR, P. N. 1987. Lead isotope evidence for the structure of the Rockall dipping-reflector passive margin. Nature, **326**, 381-383.

NAMSON, J. S. & DAVIS, J. L. 1988. Seismically active fold and thrust belt in the San Joaquin Valley, central California. Geological Society of America, Bulletin, **100**, 257-273.

NATAL'IN, B. A., PARFENOF, L. M., VRUBLEUSKY, A. A., KARSAKOV, L. P. & YUSHMANOV, V. V. 1986. Main fault systems of the Soviet Far East. In: Reading, H. G., Watterson, J. & White, S. H. (editors), Major crustal lineaments and their influence on the geological history of the continental lithosphere, 267–275, Royal Society of London.

NAYLOR, M. A., MANDL, G. & SIJPESTEIJN, C. H. K. 1986. Fault geometries in basement induced wrench faulting under different initial stress states. Journal of Structural Geology, **8**, 737–752.

NICHOLSON, C., SEEBER, L., WILLIAMS, P. & SYKES, L. R., 1986. Seismic evidence for conjugate slip and block rotation within the San Andreas Fault System, southern California. Tectonics, **5**, 629–648.

O'NIONS, R. K., HAMILTON, P. J. & HOOKER, P. J. 1983. A Nd isotope investigation of sedimentation related to crustal development in the British Isles. Earth and Planetary Science Letters, **63**, 229–240.

OXBURGH, E. R. 1972. Flake tectonics and continental collision. Nature, London, **239**, 202–204.

OXBURGH, E. R. 1974. The plain mans guide to plate tectonics. Proceedings of the Geologists Association, **85**, 299–357.

PANKHURST, R. J. 1970. The geochronology of the basic igneous complexes. Scottish Journal of Geology, **6**, 83–107.

PANKHURST, R. J. 1974. Rb–Sr whole rock chronology of Caledonian events in northeast Scotland. Geological Society of America, Bulletin, **85**, 345–350.

PANKHURST, R. J. & PIDGEON, R. T. 1976. Inherited isotope systems and the source region prehistory of early Caledonian granites in the Dalradian Series of Scotland. Earth and Planetary Science Letters, **31**, 55–68.

PANKHURST, R. J., ANDREWS, J. R., PHILLIPS, W. E. A. SANDERS, I. S. & TAYLOR, W. E. G. 1976. Journal of the Geological Society, London, **132**, 327-334. Age and structural setting of the Slieve Gamph Igneous Complex, Co. Mayo, Ireland.

PHILLIPS, W. E. A. 1973. The pre–Silurian rocks of Clare Island, Co. Mayo, Ireland and the age of the metamorphism of the Dalradian in Ireland. Journal of the Geological Society, London, **129**, 585–606.

PHILLIPS, W. E. A. 1981. The orthotectonic Caledonides. In: Holland, C. H. (editor), A Geology of Ireland, 17–39. Scottish Academic Press.

PHILLIPS, W. E. A. 1983. The Ox Mountains Inlier. In: Archer, J. B. & Ryan, P. D. (editors) Geological Guide to the Caledonides of western Ireland. Geological Survey of Ireland, Guide Series, **4**, 45–47.

PHILLIPS, W. E. A. 1983. The Callow area. In: Archer, J. B., Ryan, P. D. (editors) Geological Guide to the Caledonides of western Ireland. Geological Survey of Ireland, Guide Series, **4**, 50–52.

PHILLIPS, W. E. A. & ANDREWS, J. R. 1977. Discussion on metamorphic rocks in the southwest Ox Mountains Inlier, Ireland: their structural compartmentation and place in the Caledonian Orogen. Journal of the geological Society, London, **133**, 417–418.

PHILLIPS, W. E. A., STILLMAN, C. J. & MURPHY, T. 1976. A Caledonian plate tectonic model. Journal of the Geological Society, London, **132**, 579–609.

PHILLIPS, W. E. A., KENNEDY, M. J. & DUNLOP, G. M. 1969. Geological comparison of western Ireland and northeast Newfoundland. Memoir of the American Association of Petroleum Geologists, **12**, 194–211.

PHILLIPS, W. E. A., TAYLOR, W. E. G. & SANDERS, I. S. 1975. An analysis of the geological history of the Ox Mountains Inlier. Scientific Proceedings of the Royal Dublin Society, **5B**, 311–329.

PITCHER, W. S. 1987. Granites and yet more granites forty years on. Geologische Rundshau, **76/1**, 51–79.

PITCHER, W. S. & BUSSELL, M. A. 1977. Structural control of batholith emplacement in Peru: a review. Journal of the Geological Society, London, **133**, 249–256.

PITCHER, W. S. SHACKLETON, R. M. & WOOD, R. S. R. 1971. The Balleybofey Anticline, a solution of the general structure of parts of Donegal and Tyrone. Geological Journal, **7**, 321–328.

PLATT, J. P. 1984. Secondary cleavages in ductile shear zones. Journal of Structural Geology, 6, 439–442.

PLATT, J. P. & VISSERS, R. L. M. 1980. Extensional structures in anisotropic rocks. Journal of Structural Geology, **2**, 397–410.

PRESCOTT, W. H. & NUR, A. 1981. Accommodation of relative motion at depth on the San Andreas fault system in California. Journal of Geophysical Research, **86**, 999–1004.

PRICE, R. A. 1986. The southeastern Canadian Cordillera: thrust faulting, tectonic wedging and delamination of the lithosphere. Journal of Structural Geology, **8**, 239–254.

RAMSAY, J. G. 1967. Folding and fracturing of rocks. McGraw–Hill, New York.

RICHARDSON, S. W. & POWELL, R. 1976. Thermal causes of Dalradian metamorphism in the Central Highlands of Scotland. Scottish Journal of Geology, **12**, 237–268.

ROBERTS, J. L. 1973. The stratigraphy of the Middle and Upper Dalradian Succession between Glend Head and Greencastle, Inishowen, Co. Donegal. Proceedings of the Royal Irish Academy, **73**, 151–164.

ROBERTS, J. L. 1974. The structure of the Dalradian rocks in the Southwest Highlands of Scotland. Journal of the Geological Society, London, **30**, 93–124.

ROGERS, G., DEMPSTER, T. J., BLUCK, B. J. & TANNER, P. W. G. 1989. A high precision U–Pb age for the Ben Vuirich Granite: implications for the evolutiion of the Scottish Dalradian Supergroup. Journal of the Geological Society, London, **146**, 789–799.

RUSHTON, A. & PHILLIPS, W. E. A. 1973. A Protospongia from the Dalradian of Clare Island, County Mayo, Ireland. Palaeontology, **16**, 231–237.

RYAN, P. D. & ARCHER, J. 1977. The South Mayo trough: a possible Gulf of California type marginal basin in the west of Ireland. Canadian Journal of Earth Sciences, **14**, 2453–2461.

RYAN, P. D., SAWAL, V. K & ROWLANDS, A. S. 1983. Ophiolitic melange separates ortho and paratectonic Caledonides in western Ireland. Nature, London, **301**, 50–52.

SANDERS, I. S. DALY, J. S. & DAVIS, G. R. 1987. Later Proterozoic high pressure granulites in the northeast Ox Mountains Inlier, northwest Ireland. Journal of Metamorphic Geology, **5**, 69–85.

SANDERSON, D. J. & MARCHINI, W. R. D. 1984. Transpression. Journal of Structural Geology, **6**, 449–458.

SHACKLETON, R. M. 1958. Downward facing structures of the Highland Border. Journal of the Geological Society, London, **113**, 361–392.

SCHERMER, E. R., HOWELL, D. G. & JONES, D. L. 1984. The origin of allochthonous terranes: perspectives on the growth and shaping of continents. Earth and Planetary Science Reviews, **12**, 107–131.

SHARP, R. V. 1967. San Jacinnto fault zone in the peninsular Ranges of southern California: Geological Society of America, Bulletin, **78**, 705–730.

SHAW, H. R. 1972. Viscosities of magmatic silicate liquids: an empirical method of prediction. American Journal of Science, 272, 870–893.

SHAW, H. R. 1980. The fracture mechanisms of magma transport from the mantle to the surface. In: Hargreaves, R. (editor) Physics of magmatic processes, 2021–264. Princeton Press.

SIBSON, R. H. 1983. Continental fault structure and the shallow earthquake source. Journal of the Geological Society, London, **140**, 741–767.

SORENSON, K. 1983. Growth and dynamics of the Nordre Stromfjord shear zone. Journal of Geophysics Research, **88**, 3419–3437.

SPENCER, A. M. 1971. Late Pre–Cambrian Glaciation in Scotland. Memoir of the Geological Society of London, **6**.

SUTTON, J. & WATSON, J. V. 1955. The deposition of the Upper Dalradian rocks of the Banffshire coast. Proceedings of the Geologists Association, **66**, 101–133.

SYLVESTER, A. G. 1988. Strike slip Faults. Geological Society of America, Bulletin, **100**, 1666–1703.

SYLVESTER, A. G. & SMITH, R. R. 1976. Tectonic transpression and basement controlled deformation in the San Andreas Fault Zone, Salton Trough, California. Bulletin of the American Association of Petroleum Geologists, **60**, 2081–2102.

SYMES, R. G. 1872. One inch sheet memoirs, **number 75**, Geological Survey of Ireland.

TAYLOR, W. E. G. 1966. The structural history of the metamorphic and igneous rocks of Slieve Gamph. Unpublished Ph.D. thesis, University of Sheffield.

TAYLOR, W. E. G. 1968. The Dalradian rocks of Slieve Gamph, western Ireland. Proceedings of the Royal Irish Academy, **67B**, 63–82.

TAYLOR, W. E. G. 1969. The structural geology of the Dalradian rocks of Slieve Gamph, Counties Mayo and Sligo, western Ireland. Geologische Rundschau, **58**, 564–587.

TERRES, R.R. & SYLVESTER, A. G. 1981. Kinematic analysis of rotated fractures and blocks in simple shear. Bulletin of the Seismological Society of America, **71**, 1593–1605.

THOMPSON, R. N. 1987. Phase equilibria constraints on the genesis and magmatic evolution of oceanic basalts. Earth Science Reviews, **24**, 161–210.

TOBISCH, O. T. & PATTERSON, S. R. 1988. Analysis and interpretation of foliation in areas of progressive deformation. Journal of Structural Geology, **10**, 745–754.

TRENDALL, A. F. & ELWELL, R. W. D. 1963. The metamorphic rocks of northwest Mayo. Proceedings of the Royal Irish Academy, **62B**, 217–247.

TURNER, F. J. & VERHOOGEN, J. 1960. Igneous and metamorphic petrology. McGraw Hill. New York.

WALLACE, R. E. 1949. Structure of a portion of the San Andreas fault in southern California. Geological Society of America, Bulletin, **60**, 781–806.

WEBB, T. H. & KANAMORI, H. 1985. Earthquake focal mechanisms in the eastern Transverse Ranges and San Emigdio Mountains, southern California, and the evidence for a regional decollement. Bulletin of the Seismological Society of America, **75**, 737–757.

WELLMAN, H. W. 1955. The geology between Bruce bay and Haast River, South Westland. New Zealand Geological Survey Bulletin, **48**, (2nd edition) 46pp.

WHITE, R. S., SPENCE, G. D., FOWLER, S. R., McKENZIE, D. P., WESTBROOK, G. K. & BOWEN, A. D. 1987. Magmatism at rifted continental margins. Nature, **330**, 439–444.

WHITE, S. H. 1976. The effects of strain on the microstructures, fabric and deformation mechanisms in quartzites. Philosophical Transactions of the Royal Society of London, **A283**, 69–86.

WHITE, S. H. 1980. On mylonites in ductile shear zones. Journal of Structural Geology, **2**, 175–187.

WHITE, S. H., BRETAN, P. G. & RUTTER, E. H. 1986. Fault zone reactivation: kinematics and mechanisms. Philosophical Transactions of the Royal Society of London, A317, 81–97.

WHITE, S. H., BURROWS, S. E., CARRERAS, J., SHAW, N. D. & HUMPHREYS, F. J. 1980. On mylonites in ductile shear zones. Journal of Structural Geology, **2**, 175–187.

WILCOX, R. E., HARDING, T. P. & SEELY, D. R. 1973. Basic wrench tectonics. Bulletin of the American Association of Petroleum Geologists, **57**, 74–96.

WILLIAMS, H & HATCHER, R. D. 1982. Suspect terranes and accretionary history of the Appalachian region. Geology, **10**, 530–536.

WILLIAMS, TURNER AND GILBERT. 1982. Petrography: An introduction to the study of rocks in thin section. Freeman, San Fransisco.

WILLIS, B. 1938a. San Andreas Rift, California: Journal of Geology, **46**, 793–827.

WILSON, J. S. G. 1882. Explanation of sheet 97, northern Aberdeenshire, eastern Banffshire. Memoir of the Geological Survey.

WINCHESTER, J. A. MAX. M. D. & LONG, C. B. 1987. Trace element geochemical correlation in the reworked Dalradian metavolcanic suites of the western Ox Mountains and North Mayo Inliers, Ireland. In: Pharaoh, T. C., Beckinsale, R. D. and Rickard, D. (editors) Geochemistry and Mineralogy of Proterozoic Volcanic Suites, Special Publication of the Geological Society, London, **33**, 489–502.

WISE, D. U., DUNN, D. E., ENGELDER, J. T., GEISER, P. A., & HATCHER, R. D. 1984. Fault related rocks: suggestions for terminology. Geology, **12**, 391–394.

WOODCOCK, N. H.,& FISHER, M. 1986. Strike slip duplexes. Journal of Structural Geology, **8**, 725–735.

YARDLEY, B. W. D., LONG, C. B. & MAX, M. D. 1979. Patterns of metamorphism in the Ox Mountains and adjacent parts of western Ireland. In: Harris, A. L., Holland, C. H. & Leake, B. E. (editors), The Caledonides of the British Isles reviewed, Special Publication of the Geological Society, London, number **8**, 369-75.

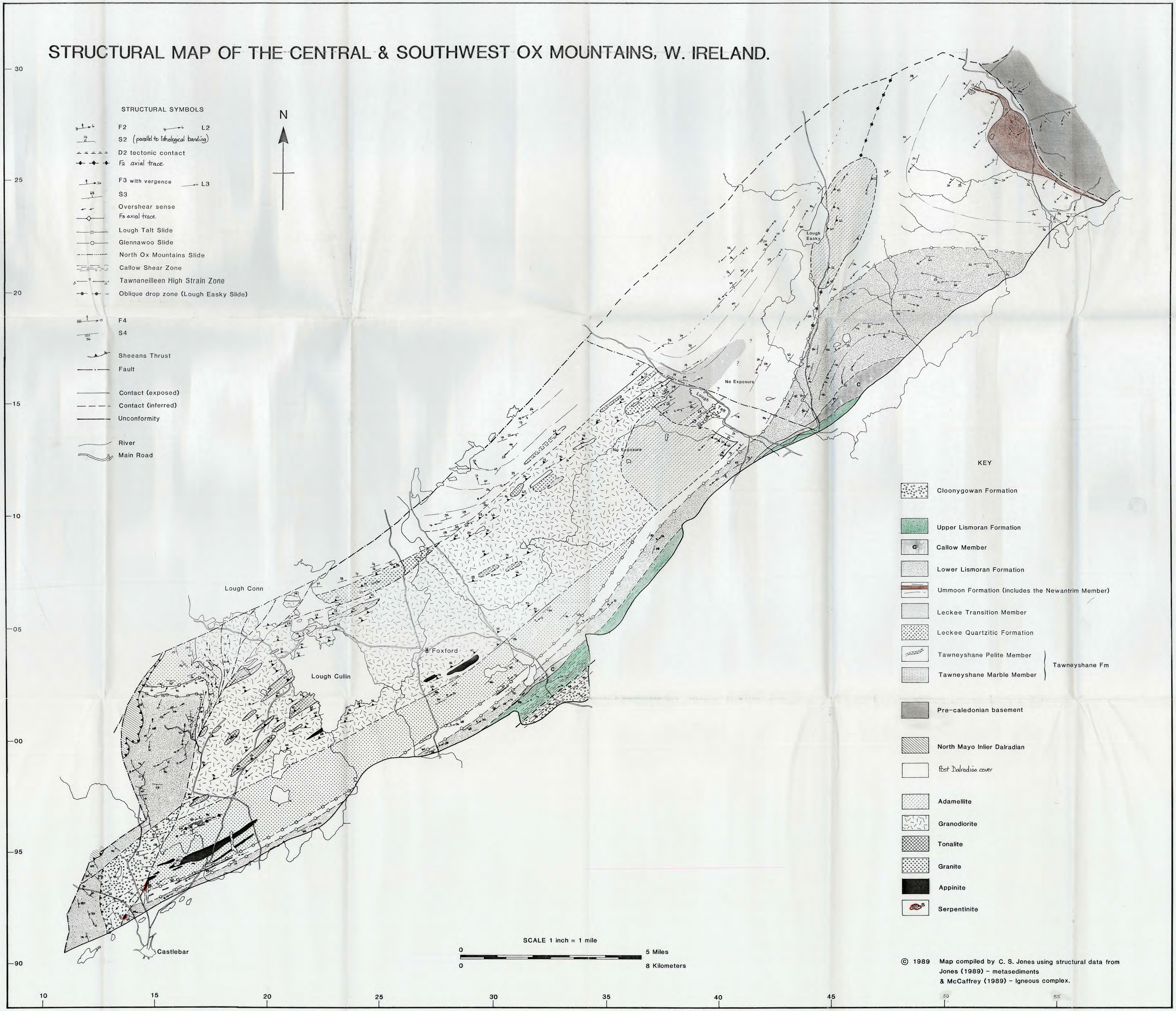
YARDLEY, B. W. D. 1980. Metamorphism and orogeny in the Irish Dalradian. Journal of the Geological Society, London, **137**, 303–309.

YARDLEY, B. W. D. & LONG, C. B. 1981. Contact metamorphism and fluid movement around the Easky adamellite, Ox Mountains, Ireland. Mineralogical Magazine, **44**, 125–131.

YEATS, R. S. 1981. Rifting and rafting in the southern California borderland. In: Dickinson, W. R. & Grantz, A., (editors) Proceedings of conference on Geological Problems of the San Andreas Fault System. Stanford University Publications in the Geological Sciences, **7**, 307–322.

FIG 2.4

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



© 1989 Map compiled by C. S. Jones using structural data from
Jones (1989) – metasediments
& McCaffrey (1989) – Igneous complex.