GENESIS AND STRUCTURAL RELATIONS OF MOINE MIGMATITES

VOLUME 1: TEXT

Thesis submitted in accordance with the requirements of the University of Liverpool for the degree of Doctor in Philosophy by David Barr.

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SUMMARY

The problem of the origin of the Moine migmatites was tackled by means of detailed mapping at 11 key areas in the Northern Highlands of Scotland. The timing of migmatisation was defined relative to tectonic fabrics in the rocks, particularly those related to the Sgurr Beag Slide, a major early Caledonian thrust within the Moines, which is associated with intense ductile fabrics in the wall rocks.

At Mallaig, in the non-migmatitic western Morar Division, a simple structural sequence associated with garnet-grade metamorphism was defined. At Inverie, 10km to the east, similar rocks are in tectonic contact with kyanite-grade Morar Division migmatites across the Knoydart Slide; this "early, trondhjemitic migmatisation" predates sliding, and formed quartz-oligoclase lits in pelitic gneisses. The Sgurr Beag Slide separates the Morar and Glenfinnan divisions of the Moines, and at Acharacle, Kinlochourn and Ben Klibreck forms the western limit to this early migmatisation, which is ubiquitous in the Glenfinnan Division. Geothermometry indicates a jump in temperatures of 40-50°C across the slide. Isolated outcrops of migmatitic Glenfinnan Division lithologies, probably resting on the slide, occur within the Morar Division of the Ross of Mull and Sguman Coinntich. Loch Quoich and Glenfinnan lie on the boundary between the Glenfinnan and Loch Eil divisions (the Loch Quoich Line), and rocks of both divisions contain bodies of granitic gneiss - this was considered to represent an early or pre-tectonic granite. The Loch Quoich Line is not a slide or unconformity, but is a zone across which the latest, upright, post-slide folds tighten. "Late migmatisation" (post-sliding) is widespread in the eastern areas, and is represented by pegmatites, local segregations and early Caledonian intrusions. One such pegmatite complex, at Lochan Coire Shubh, was mapped in detail. Contact migmatites in the inner aureole of the forceful Strontian granite were considered to represent deformed regional migmatites, rather than new segregations.

The trondhjemitic migmatites occur in the kyanite and sillimanite+muscovite zones, and the K-feldspar-free lits are inconsistent with partial melting of quartz+plagioclase+muscovite-bearing pelites. P,T estimates (obtained using published geothermometers and new geobarometers) of 630-650°C at 6-6.5kb lie below those required for partial melting of K-feldspar-free pelites. Based on a statistical survey of the bulk compositions of migmatites and non-migmatites within the migmatite complex, it was concluded that migmatisation was favoured by a high plagioclase content and inhibited by a high mica content - K-feldspar content, and the Na/Ca ratio, appear to be unimportant, again ruling out partial melting. However, some unusual late segregations in the Quoich granitic gneiss and nearby metasediments do appear to be partial melts, while the pegmatites of the Lochan Coire Shubh complex could be partial melts of deeper Moine rocks.

Regional correlation and consideration of published radiometric dates suggests that the Sgurr Beag Slide is a major early Caledonian thrust, which foreshortens a Precambrian (Grenville?) metamorphic complex which contained essentially recumbent folds and in which metamorphic grade rose to the east, culminating in migmatisation in the Glenfinnan Division and eastern Morar Division; metamorphic grade decreased upwards in the Glenfinnan Division, and the stratigraphically-younger Loch Eil Division, although metamorphosed in the same event, is often lower-grade and non-migmatitic. Caledonian metamorphic grade also rose to the east, with local hot spots, and the peak of metamorphism postdated sliding and subsequent folding, so that isograds cross the folded Sgurr Beag Slide at Acharacle.

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CHAPTER 1

Introduction and aims

1.1 Aims of the project

The aims of the project were twofold: to determine the mechanisms by which the regional migmatites of the Northern Highlands were formed, and to date their formation relative to structural events in the migmatites and surrounding rocks.

The main priority was the origin of the regional "lit-par-lit" migmatites - the "trondhjemitic migmatites" or "permeation gneiss" of the Geological Survey and earlier authors (cf. Johnstone 1975, Read 1931), which occupy much of the eastern part of the Moine outcrop and are distinguished by their K-feldspar free leucosomes. Some attempts were also made to characterise the late cross-cutting pegmatites, which are quite distinct from the regional migmatites, and the West Highland granitic gneiss. which was considered by some (e.g Dalziel 1966) to be a migmatite.

Dating of the migmatites was carried out using conventional methods of structural analysis, with deformation of leucosome material being taken to indicate that migmatisation predated that tectonic phase. Since the problem to be solved was a regional one, it was approached by detailed mapping at a number of critical areas (selected from the literature and in discussion with A.L. Harris), which were large enough to determine the structural relations of the migmatites and to define a local structural sequence which could be fitted into those of other workers. Of particular importance was the Sgurr Beag Slide (Tanner et al. 1970, Tanner 1971, Rathbone & Harris 1979, Rathbone 1980), a major structural feature (probably an early thrust) which separates the Morar and Glenfinnan divisions of the Moines (Johnstone 1975) and often forms the western boundary to regional migmatisation. The strong fabric associated with the slide facilitates structural dating of the migmatites and links the structural sequences in the two divisions. A major part of the work involved correlating these sequences, and relating them to existing structural and radiometric dates.

The question of the origin of the migmatites was approached from two directions. Temperatures and pressures obtained by geothermometry and geobarometry were compared with grids of likely pelite melting reactions, and predictions about the melt compared with petrographic observations. A number of new analyses were performed, which with others from the literature were used to determine the factors which controlled migmatisation (it having become apparent that within a given area, some lithologies were more prone to migmatisation than others). Large-scale metasomatism was not specifically tested for, it having been ruled out by Steveson (1968, 1971), but no chemical features were found in the migmatites which could not be matched with those of low-grade rocks elsewhere in the Moines. Detailed analyses were performed on specific later migmatites which did not belong to the regional trondhjemitic suite and appeared to have undergone partial melting.

1.2 Structure of the thesis

The thesis falls into two main sections - Chapters 2-14 are concerned with the field relationships of the migmatites and the structure of the Moines, while Chapters 15-17 are concerned with the origin of the migmatites. Chapter 18 briefly summarises the main conclusions.

Chapter 2 incorporates a summary of the regional geology of the Moines, and sections defining the structural and geochemical terms used.

Chapters 3-13 each cover one field-area (locations on Fig. 1.1) and contain descriptions of lithologies, structure, metamorphism and migmatisation as appropriate. Mallaig lies in the garnet-grade, non-migmatitic Morar Division where structures are relatively open, and was chosen to provide a basic structural sequence, parts of which might be recognised elsewhere. At Inverie, a slide separates the migmatitic Knoydart Pelite from Morar Division rocks similar to those at Mallaig. Kinlochourn and Acharacle lie on the main trace of the Sgurr Beag Slide, where later deformation and metamorphism are more intense than in the west. On the Ross of Mull, where Glenfinnan Division rocks ocupy a synform and rest on a slide, later deformation and metamorphism are both very slight and some estimates can be made of strain in the slide zone. At Sguman Coinntich, where migmatitic Glenfinnan Division pelites are enclosed by a slide, regional structures are moderately SE-dipping and resemble those of the Moine Thrust Zone. On Ben Klibreck, a major slide is present along with prominent Lewisian (basement) inliers and later granitic intrusions; the structural sequence of the Moine Thrust Zone can be recognised in the west. At Lochan Coire Shubh, a late pegmatite complex is present within the Glenfinnan Division. The Loch Quoich and Glenfinnan sections both lie on the Loch Quoich Line (Clifford 1958a), at the boundary between the Glenfinnan and overlying Loch Eil divisions, and contain representatives of the West Highland granitic gneiss. Strontian is an example of a forceful Newer Granite (Read 1961); Ashworth & Chinner (1978) claimed that pelites in the aureole have undergone partial melting. A sample from Loch Ruthven, south of the Great Glen, was included in the P,T study discussed below.

Chapter 14 summarises these observations, and contains a major section on the nature of the Sgurr Beag Slide; the slide is then used as a structural marker, and the sequence so derived correlated with published radiometric dates.

In Chapter 15, garnet-biotite temperatures are combined with published geobarometers and some newly derived for this thesis to obtain P,T estimates for the various field-areas. These are compared with grids of melting reactions (taken mainly from the literature) to determine if partial melting is a likely mechanism for the formation of the regional migmatites.

Chapter 16 contains a general section on the chemistry of the Moines (emphasising how their compositions permit only a limited range of metamorphic index minerals) but is mainly concerned with the origin of the migmatites. Rocks from the same localities are split into migmatites and non-migmatites and their chemical differences (determined by statistical tests and from triangular diagrams) used to infer chemical controls and hence likely mechanisms of migmatisation. A final section investigates possible examples of partial melt migmatites by major and trace element modelling (including rare earths).

Chapter 17 presents the conclusions on migmatisation mechanisms, and includes some additional petrography - sections are included on the regional trondhjemitic migmatites, an unusual K-feldspar-bearing migmatite, the granitic gneiss and the late cross-cutting pegmatites.

Chapter 18 briefly summarises the main conclusions, referring back to earlier chapters for additional detail, and presents some regional tectonic models and cross-sections.

Microprobe (mineral) and XRF (whole-rock) analyses are presented in the Appendices, along with the sources for analyses taken from the literature and a list of sample numbers and locations.

1.3 Summary of previous work

Only a few general papers will be mentioned here. Section 2.1 comprises a summary of the regional geology of the Moines, and other papers of local relevance will be referred to in the appropriate chapters.

The most important previous work is that related to the Sgurr Beag Slide. Tanner (1971) defined the slide at Kinlochourn, and recognised that the successions on either side (the Morar and Glenfinnan divisions) were quite distinct and could not be correlated; Tanner _et al._ (1970) extended this interpretation to other parts of the Moines. Rathbone (1980 - also Rathbone & Harris 1979) visited several outcrops of the slide and recognised a characteristic strain pattern in the adjacent rocks; this meant that the slide could be identified where the stratigraphic evidence was less clear-cut, and he was able to extend it beyond the area covered by Tanner _et al._ (1970). Regional structural correlations have been suggested by Tobisch _et al._ (1970) and Powell (1974); these will be discussed in section 14.3.

Early work on the migmatites by Read (1931) and Cheng (1944) in Sutherland and Harry (1952, 1953) and Dalziel (1966) in Inverness-shire was largely field-based and emphasised metasomatism or igneous injection, although the few analyses used were not statistically significant. In addition, the Sutherland migmatites are complicated by the presence of many Lewisian inliers and of Caledonian intrusions, while in Harry's and Dalziel's areas, numerous Caledonian pegmatites occur which were not distinguished from the leucosome of the Precambrian pelitic gneiss. Steveson (1968, 1971) carried out a detailed statistical comparison of the high-grade and low-grade Morar Pelite and found no evidence for metasomatism. Brown (1967) and Soper & Brown (1971) emphasised the role of Na-metasomatism in producing the Loch Coire migmatite complex; they did not distinguish between regional migmatitic _lits_ and later intrusions, and also failed to recognise that a major slide juxtaposes the migmatites and the non-migmatites, which belong to separate divisions and would be expected to have different original compositions. The present work is concerned mainly with the origin of the leucosomes in the pelitic gneiss, a question which had previously only been addressed by Steveson.

CHAPTER 2

Concepts, definitions and introduction to the Moines

2.1 Regional geology of the Moines

This section is based on existing published work, with comments where it conflicts with the conclusions of this thesis. Several recent summaries exist: Johnstone *et al.* (1969) covered all aspects of the Moines, Johnstone (1975) updated the stratigraphy, Harris *et al.* (1978a) included a brief section on the Moines, while several papers in Harris *et al.* (1979) are relevant and include most of the recent references (especially Watson & Dunning, pp. 67-75, Johnson *et al.*, pp. 165-169, and Fettes, pp. 307-321). A summary of the older literature can be found in Phemister (1960), pp. 17-40.

According to Johnstone (1975), the Moines are "a succession of metasedimentary rocks of Scotland and Ireland which, lying east or southeast of the Caledonian Front thrust belt, unconformably overlies the Lewisian Complex and is older than the Dalradian Supergroup". Only the Scottish segment will be considered here (Fig. 2.1; the location of the Ross of Mull inlier, absent from Fig. 2.1, is shown in Fig. 2.2).

The Lewisian Complex consists mainly of tonalitic gneisses and granulites, severely metamorphosed at *c.* 2700Ma B.P. (Early Scourian or Badcallian), and variably reworked from 2600 to 1700Ma B.P. (Late Scourian to Laxfordian) - e.g. Watson & Dunning (1979). Most Scourian areas consist of high-pressure granulites, severely depleted in incompatible elements (Tarney *et al.* 1972, 1979), while the Laxfordian areas are mainly in amphibolite facies and are often affected by granite intrusion and migmatisation. Minor metasediments and later intrusions are present within the complex.

Lewisian rocks and their cover of sediments form the foreland to the Caledonian orogen in Scotland, and the former also occur as inliers within the Moines. The Glenelg inliers (Fig. 2.1) preserve much of their Lewisian structural and metamorphic character (Barber

& May 1976), and basal Moine conglomerates have been reported from their margins (Peach et al. 1910, Ramsay 1958b). The westernmost inlier preserves K/Ar mineral ages of 2200-1600Ma (Moorbath & Taylor 1974) - possibly partially reset Scourian ages. Elsewhere, Lewisian inliers are usually identified by their lithological character (tonalitic or hornblendic gneisses or granulites, basic intrusions not found in the adjacent Moines, distinctive metasediments such as graphitic schists and marbles). The presence of hornblende is often considered diagnostic, since within the Moines, this mineral is normally restricted to readily identifiable basic meta-igneous rocks and calc-silicates.

In Morar, Lewisian inliers form the cores of early fold-nappes within the Moines (e.g. Powell 1974), but their status in Ross-shire and Sutherland was questioned when it was found that the Moine metasediments younged towards them (Sutton & Watson 1953, 1954, 1955; Ramsay 1958a). However, it was later accepted that these inliers had been thrust into their present position (inserted into the Moine sequence, according to Sutton & Watson 1962). Tanner et al. (1970) showed that Sutton & Watson's correlation of the Morar Pelite with the pelite overlying the Lewisian was invalid, and suggested that the inliers lay at the base of the Glenfinnan Division (see below). Their Scourian age has since been confirmed by Rb/Sr whole-rock dating (Moorbath & Taylor 1974).

The Dalradian (Harris & Pitcher 1975, Harris et al. 1978b) is a thick series of geosynclinal sediments, ranging in age from late Precambrian (c. 700Ma) to Cambrian, and deformed and metamorphosed in the 490-500Ma Grampian event (Fettes 1979).

The "Young Moines" or Central Highland Granulites (Johnstone 1975 - so-called for their granulitic appearance, not for their metamorphic grade) occupy much of the Grampian Highlands, apparently pass conformably into the Dalradian (Treagus & King 1978) and are generally considered to have suffered only Caledonian (Grampian) deformation and metamorphism. These form the Grampian Group of Harris et al. (1978b) or the Grampian Division of Piasecki (1980). The latter author considers that these rocks post-date metamorphism in the "Old Moines", but have suffered some deformation and metamorphism during the Morarian event (see below), and may not pass stratigraphically into the Dalradian.

North of the Great Glen, the Moines appear to have suffered Precambrian deformation and metamorphism (see below), and it is these "Old Moines" which form the subject of this thesis. Several inliers of "Old Moines" have recently been mapped south of the Great Glen, by Piasecki (1980) on Speyside (his Central Highland Division) and by A.J. Highton & A.L. Harris (verb. comm.) in the area around Loch Ruthven; the latter area was briefly visited in the course of the present study, and samples from there were included in the pressure/temperature determinations of Chapter 15. On Speyside, the "Young Moines" apparently overlie the "Old Moines" on a tectonised unconformity. At Loch Ruthven, low-grade weakly deformed psammites ("Young Moines" or Morar Division "Old Moines") young towards overlying complexly and strongly deformed migmatites of Glenfinnan Division aspect, and are separated from them by a major slide.

Johnstone et al. (1969) divided the Moines north of the Great Glen Fault into the Morar, Glenfinnan and Loch Eil divisions (Fig. 2.1, Table 2.1). The stratigraphy of the Morar Division is best-known south of Glenelg, and several local successions have been widely correlated (Table 2.2). Further north, where the Morar Division is largely psammitic, highly deformed and cut by numerous slides (Mendum 1979), detailed stratigraphies have not been erected.

The Morar Division is separated from the Glenfinnan Division by the Sgurr Beag Slide, which is probably a major early Caledonian thrust with tens of kilometres displacement (section 14.2; Tanner et al. 1970, Rathbone 1980). It can be traced at least from the Loch Coire area to Ardnamurchan (Fig. 2.1), and may be brought back down by later folding to limit the Glenfinnan Division outlier on Mull (Chapter 7).

The Glenfinnan Division is more pelitic than the Morar Division (Table 2.1), the rocks are usually highly-deformed, and suitable lithologies are migmatitic. Local lithostratigraphic successions cannot be widely correlated (cf. the differences between Argyll and Western Inverness-shire, Table 2.2). Given the large displacements likely on the Sgurr Beag Slide, the Glenfinnan Division cannot be assumed stratigraphically to overlie the Morar Division, and indeed, given the common occurrence of Lewisian inliers near its base, associated with a distinctive feldspathic psammite (the Reidh Psammite - Tanner 1971), it is more likely to be a lateral equivalent, perhaps from a separate sub-basin (Tanner et al. 1970, Johnstone et al. 1979, Rathbone 1980). The Sutherland migmatites, at least in the vicinity of Loch Coire (Chapter 9), are probably an extension of the Glenfinnan Division, the wider outcrop being due in part to generally lower dips. In the Strath Halladale area, probable Glenfinnan Division rocks are intruded by large volumes of late Precambrian granite (Johnstone et al. 1979, Lintern et al. in press).

The Loch Eil Division is largely psammitic (Tables 2.1, 2.2) and structurally overlies the Glenfinnan Division (Chapter 11; Dalziel 1966). The boundary between the two has been called the Loch Quoich Line (Clifford 1958a) and interpreted as a slide (Clifford op. cit., Johnstone 1975) or an unconformity (Lambert et al. 1979, Winchester et al. 1981). In fact there is a normal stratigraphic passage from the Glenfinnan Division into at least the lower part of the Loch Eil Division (Chapter 11; Roberts & Harris in press - cf. Strachan 1982), and they share the same extensive structural history, as well as geochemical features which distinguish them from the Morar Division (Johnstone et al. 1979). Extreme ductile strains similar to those encountered at the Sgurr Beag Slide (section 14.2; Rathbone & Harris 1979, Rathbone 1980) are absent, although moderate strains do occur in a zone which runs close to the boundary between the divisions, and corresponds to a late upright fold limb (Chapter 11). This forms the boundary between a "steep belt" in which late upright folds are tight and steeply-plunging, and a "flat belt" in which they are open and gently-plunging; in detail, it deviates from the lithostratigraphic boundary.

Spatially associated with the Loch Quoich Line from Glen Affric southwards is a series of distinctive granitic gneiss bodies

(Fig. 2.1). These have been interpreted as migmatised Moine pelites and semipelites which have been subjected to K-metasomatism in a major shear zone (Dalziel 1966), but are more convincingly explained as a suite of pre-tectonic granite intrusions (section 17.3.1).

The existence of a Precambrian metamorphic event affecting the Moines has already been alluded to, and an appraisal of published dates will be given in section 14.4. Structures and a metamorphic fabric in the adjacent Moines predate the 560Ma Carn Chuinneag granite (Wilson & Shepherd 1979), and a large number of 700-750Ma dates have been obtained from early pegmatites, using a variety of methods (Long & Lambert 1963, Lambert 1969, Pidgeon & Johnson 1973, van Breemen et al. 1974). These dates correspond to the Morarian event of Lambert (1969) (the alternative term Knoydartian (Bowes 1968, 1980) has never gained acceptance in the Moine literature). Convincing Grenvillian (c. 1000Ma) dates have been obtained from the granitic gneiss (Brook et al. 1976, Aftalion & van Breemen 1980), but the c. 1000Ma dates from Moine pelites (Brook et al. 1977, Brewer et al. 1979) are less convincing. Numerous c. 450Ma dates on pegmatites, minor intrusions and metamorphic minerals confirm the importance of Caledonian overprinting (e.g. Long & Lambert 1963, van Breemen et al. 1974, Brewer et al. 1979).

Several interpretations have been advanced to explain these dates. The Moines may have been involved in a major Grenvillian orogeny (Brewer et al. 1979), with the Morarian pegmatites either giving partially-reset Grenville ages, or having formed in a relatively minor thermal event. Alternatively, the Morarian may have been the main orogeny, with the old pelite dates recording diagenetic or source-rock influences, the granitic gneiss possibly being a basement slice (Lambert et al. 1979). Piasecki (1980), working SE of the Great Glen, prefers significant tectonometamorphic activity at all three stages. In any case, it is likely that the early migmatisation and metamorphic fabric in the Moines are Precambrian, while structures such as the major slides and upright NNE-SSW folds, and the later overprinting metamorphism and migmatisation, are Caledonian (cf. Chapter 14).

The structure of the Moines of the Northern Highlands has been extensively studied, with several correlations proposed (e.g. Tobisch et al. 1970, Powell 1974, Brown et al. 1970, Soper & Wilkinson 1975). Some phases of folding may be of only local significance, and the correlation of others is open to question; the presence of slides of status equal to or greater than the folding (e.g. Rathbone & Harris 1979, Rathbone 1980, Mendum 1979) has particularly profound implications for correlations across strike. One major dilemma is the apparent correlation of the Caledonian D2 structures of Sutherland (Soper & Brown 1971, Soper & Wilkinson 1975) with the pre-Caledonian structures of Morar (Powell 1974). An attempt to resolve this, and other problems of regional correlation, will be made in Chapter 14. Only a brief summary of regional structure is given here. North and west of the vicinity of the Strathconon fault, axial planes dip southeast at moderate angles, and folds tend to be overturned to the west, and have strongly curvilinear axes around a SE-plunging extension lineation. The major folds are Caledonian, but an earlier fabric is preserved. To the south, pre-Caledonian recumbent folds are cut by the Sgurr Beag Slide and related early Caledonian thrusts, and are refolded by upright N-S to NNE-SSW - trending structures, which are at their tightest in the Glenfinnan Division.

Work on the Moine Thrust zone is summarised in McClay & Coward (1981). The relevant features are that it is a basement-involved thrust zone (Lewisian occurring in imbricates and intermediate nappes) in which an early phase of mylonitisation (correlated by Soper & Wilkinson 1975 with D2, and less convincingly with D1, in the Moines) was succeeded by brittle thrusting and open folding. The Cambro-Ordovician platform sequence (extending at least to the Arenig - Higgins 1967) is involved, and broadly contemporaneous alkaline intrusions are dated at 430±4Ma (van Breemen et al. 1979a), placing contraints on the timing of Caledonian deformation in at least the western part of the Moine Nappe.

The metamorphic grade of the Moines has been the subject of some discussion since Kennedy (1949) defined isograds in the Morar area, based on calc-silicate assemblages; a summary is given in

Fettes (1979). Clearly, given the polymetamorphic nature of the Moines, and the scarcity of good indicators of metamorphic grade (especially pelitic index minerals), any overall description must be very generalised.

Winchester's (1974) map is reproduced as Fig. 2.2, although it is known to be incorrect in certain areas. It was produced using published occurrences of pelitic index minerals, and an empirical correlation of calc-silicate assemblages with Barrovian isograds in the Central Highlands. Winchester believed that the CaO/Al_2O_3 ratio in a calc-silicate was important in determining the grade at which a particular index mineral (e.g. hornblende, clinopyroxene, bytownite) appeared, so he included a correction for bulk composition in his calc-silicate/pelite correlation. Clearly this can only be a rough correction (e.g. the K_2O content must be relevant to biotite-hornblende reactions), and in any case it is the mineral composition (only indirectly related to the rock composition) which is important in defining the P,T range of a particular assemblage (cf. Winchester 1981 and Wells 1981). Calc-silicate and pelite reactions will have different slopes in P,T space (e.g. Wells 1981), and different responses to variations in X_{H_2O} and X_{CO_2}. The significance of the pressure-sensitive kyanite/sillimanite isograd is particularly debatable. Although geobarometry gives broadly similar pressure estimates throughout the Northern Highland Moines (c. 6kb - section 15.3.2), these are very different from those obtained by Wells & Richardson (1979) in the Central Highlands (9-13kb), and it was in this latter area that Winchester (1974) calibrated his calc-silicate assemblages against pelitic index minerals. In addition, Wells (1979) has suggested that most of the sillimanite occurrences used by Winchester (op. cit.) were related to thermal aureoles, or to late-stage decompression on uplift, and should not be correlated with the regional metamorphic assemblages.

Winchester (1974) claimed that his map recorded the "high water mark" of metamorphism, whether Caledonian or Precambrian, and that retrogression could always be recognised. This is certainly not true of the area around Loch Quoich, where Precambrian high-grade gneisses which locally have suffered later partial melting are placed in the garnet zone (Chapter 11, sections 17.2, 17.3.2). The retrogression in this area has also been noted by Tanner (1976) and relic high-grade assemblages survive in some metabasic rocks, and less frequently in calc-silicates. In fact, Winchester only recognised retrogression in the west, where it was accompanied by strong deformation. Thus while most of the kyanite/sillimanite areas can safely be regarded as high grade, some of the lower-grade areas may be retrogressive, and the temperature significance of the kyanite/sillimanite distinction is questionable.

The most striking feature of the map is the coincidence of the high-grade zone (and of the regional migmatites) with the Glenfinnan Division, reflecting disruption of a Precambrian metamorphic complex by the Sgurr Beag Slide. Within the Morar Division, metamorphic grade increases eastwards. In Sutherland, this may be tectonic, resulting from stacking of successively higher-grade (more easterly, and deeper) crustal slices on Caledonian slides (cf. Mendum 1979), but there is also a contribution from a regional eastwards increase in Caledonian grade (e.g. severe grain coarsening on Ben Klibreck post-dates all tectonic fabrics - section 9.3). In the south, the calc-silicates largely record Caledonian metamorphic grade (Tanner 1976, Fettes 1979), but Lambert et al. (1979) suggest that the biotite-grade enclave in Morar is due to an area of Caledonian retrogression (which was more effective at depth) being brought up in the core of the Morar Antiform, the surrounding garnet zone being Precambrian.

It is important to note that although calc-silicate minerals may record the Caledonian event, this does not imply that grosser textural features (in particular migmatitic banding) were reworked. In fact it is concluded in this thesis that the form of Precambrian migmatitic veins and lits was not, in general, altered by even high-grade Caledonian metamorphism, although individual minerals may have recrystallised and re-equilibrated. From the results of Chapter 15, it can be concluded that suitably-selected garnets in pelitic gneisses can preserve their original compositions (and give reliable Precambrian P,T estimates), even where the calc-silicates have been completely "reset" by Caledonian metamorphism. Thus geothermometry may ultimately offer a better means of defining the Precambrian metamorphic zones.

The Moines are intruded by several suites of Caledonian igneous rocks. The Carn Chuinneag complex (Fig. 2.1) was intruded c. 560Ma ago (Pidgeon & Johnson 1974), and consists dominantly of deformed potassic granite, although compositions ranging from gabbro to riebeckite granite occur (Wilson & Shepherd 1979). Later intrusions such as the Strontian complex are more typically calc-alkaline (Brown 1979, Pankhurst 1979), with low initial $^{87}Sr/^{86}Sr$ ratios and dates ranging from 440-390Ma (Pidgeon & Aftalion 1978, Halliday et al. 1979). The distinctive Assynt suite of alkaline intrusions occurs south of Loch Erribol (Parsons 1979), and has been dated at c. 430Ma (van Breemen et al. 1979a), and a 456±5Ma deformed syenite occurs at Glen Dessary, 10km east of Loch Morar (Richardson 1968, van Breemen et al. 1979b). Several suites of minor intrusions exist, but tend to be localised in particular areas (Smith 1979, Fettes & MacDonald 1978).

2.2 Terminology

2.2.1 Structural principles and terminology

i) Mesoscopic structures

Orientation Fig. 2.3(a) shows the system of orientation measurement used for planes and lines. Note particularly the distinction between pitch and plunge - a fold axis pitching at 90° in a gently-dipping axial plane would have a plunge of only a few degrees.

Stereograms All stereographic projections are lower hemisphere plots on an equal-area (Schmidt) net. Orientations are relative to grid north, and planes are represented by the projections of their poles. Contouring was done by hand, using the method given in Phillips (1971, pp. 64-65).

Structural elements These are defined by letters: S for planar structures (e.g. axial planes, cleavages), L for lineations, F for folds or fold axes, and D for a general deformation event. S and L are also used to describe the degree of planarity or linearity of a fabric; an LS fabric has some elements of both (e.g. a shape fabric produced by simple shear, with axial ratios 4:2:1).

Fabrics Slaty cleavage is a penetrative fabric defined by orientated mica flakes; schistosity is a (generally higher-grade) penetrative fabric in which coarse aligned micas alternate with thin (few millimetres) lenses and ribbons of quartzofeldspathic material. Gneissosity is a coarser fabric with segregation of micaceous and quartzofeldspathic material apparent in hand specimen - quartzofeldspathic bands or lenses will usually be several millimetres to a few centimetres thick, and coarse-grained.

A crenulation cleavage is produced by small-scale (*c.* 1mm) buckling of schistosity, with development of new orientated mica in the long limbs of the buckles; where a distinct cleavage has not formed, a notional S-surface can be defined by the axial planes of the crenulations. In strain-slip cleavage, the new mica-filled cleavage planes are more coarsely spaced (several millimetres) and the old cleavage in intervening microlithons shows little or no rotation.

A grain or grain-aggregate shape fabric is defined by the systematic orientation of the principal axes of grains or aggregates of grains. A location fabric is defined by alternating layers of different composition, and need not be associated with any crystallographic or grain-shape orientation. Transposition describes the breaking up of elements of an early fabric by a later one, and their rotation into parallelism with the latter. This is particularly important in deformed migmatites, where intermediate stages in the transposition process are commonly preserved.

Linear fabrics can be defined by fold or crenulation axes, intersections between S-surfaces (e.g. cleavage/bedding), necking points of boudins, etc. Mullions are rib-like linear features produced, for instance, by low-angle cleavage/bedding intersections. Mineral lineations can be defined by alignment of the long axes of mica plates, hornblende prisms, etc. (possibly related to growth in a deviatoric stress field, or bodily rotation of inequant grains during progressive strain) or by the shape of normally-equant mineral grains or grain-aggregates (usually quartz, possibly deformed pebbles). This last type can be an indicator of the finite strain suffered by the rock.

Folds The terminology used to describe fold orientation and tightness is summarised in Figs. 2.3 and 2.4. Note the distinctions between the hinge (the point of greatest curvature) and the crest of a fold, and between the axial trace (intersection of the axial plane with the ground surface), the strike of the axial plane, and the plunge of the hinge. There has been some confusion over these points in the Moine literature (e.g. Clifford 1958a, McIntyre 1952). Where axial planes dip gently SE (as in the northern Moines) reclined folds with SE-plunging axes will have axial traces which trend NE-SW - i.e. it is the dip and strike of the axial plane which is of primary importance in relating minor folds to outcrop patterns, not the plunge of the axis. Fold axes in the Moines are often strongly curvilinear, but lie within a constant axial plane. Where these curving axes result in a flattened "test-tube" form, the structure is described as a sheath fold (Quinquis et al. 1978). The distinction between anticline/ syncline and antiform/synform is made clear in Fig. 2.4: the former is determined by the presence of older of younger rocks in the core, the latter solely by whether the structure opens upwards or downwards.

Terms used to describe fold styles are given in Figs. 2.5 and 2.6 - note that "asymmetric" is independent of the orientation of the axial plane. A more precise classification is given in Ramsay (1967, p.365), but as this requires detailed measurements to be made on the profile plane of the fold, it is unsuitable for use in the field. These classes are given in Fig. 2.7, but where used, have only been determined by a qualitative assessment of the field appearance. Fold interference patterns are described by the classification of Ramsay (1967, pp. 520-533), based on two interfering sets of ideal similar folds (Fig. 2.7). Natural examples deviate somewhat from these patterns, and some (particularly Type 1) can be produced by a single episode of deformation.

ii) Regional structures

Large-scale folds are described using the same terms as small-scale examples.

Vergence is used to describe the asymmetry of congruent minor folds (Fig. 2.8(a)), not the direction of overturning, which can only be applied to alternate limbs, and in any case is implicit in the statement of axial plane orientation. To avoid ambiguity, the nature of the major structure will always be quoted - e.g. at A on Fig. 2.8(a), the minor folds are verging towards an antiform lying to the east (or a synform to the west), rather than simply "easterly-verging", while those at B verge towards an antiform lying to the west. Vergence can also be obtained from the relationship of axial planar cleavage to bedding.

Younging directions (determined from sedimentary structures) are taken to be at right angles to bedding and shown on maps or sections as a "Y", used as a headless arrow pointing in the younging direction. Facing (Fig. 2.8(b)) is the direction in which younger beds come on within the axial plane of a fold, or within a tectonic fabric (cf. Shackleton 1958). This is generally taken to be at right angles to the cleavage/bedding intersection or fold axis, but in areas of high strain where these are curvilinear it can more usefully be defined parallel to the extension lineation or quartz shape fabric. Facing thus defined is a vector, and could be quoted as a lineation - e.g. upward-facing along 25/135 - in general this can adequately be expressed as "upward-facing to the northwest". Recumbent structures can be described as east-facing, west-facing, etc., and upright structures as upward or downward-facing. Note that in an area of multiple deformation, different generations of structure may have different facing directions. An extension of the concept of facing has been applied to the Sgurr Beag Slide. Assuming that it was originally a gently-dipping thrust, and the Glenfinnan Division overlay the adjacent Morar Division, then subsequent structures can be described in terms of their "facing" across the slide. This usage will be distinguished by quotation marks.

The term thrust will be used for a low-angle (<30°) reverse fault or shear zone, a low-angle fault zone comprising ramps (steep sections) and flats (nearly bedding-parallel sections), or a fault or shear zone which it is believed originally had such an orientation. Definitions based on shortening of markers (e.g. McClay 1981) are potentially ambigous in already-folded areas where bedding was probably not horizontal, and where subsequent tilting complicates recognition of an alternative horizontal marker. If a thrust is permitted to cut down-section (e.g. in beds which dip more steeply than the thrust plane - cf. Allmendinger 1981) then bedding will be extended - Fig. 2.8(c). A nappe is a body of rock bounded by major thrusts, generally tens of kilometres in length, or defined by a major recumbent fold. A duplex is a body of rock bounded by floor and roof thrusts.

iii) Deformation histories

Successive phases of deformation are described by numbers (S1, S2, F2, D3, etc.). Subscripts indicate that this sequence only applies to a particular area e.g. $S1_a$, $F2_q$. The local sequences have been established in small areas (a few square kilometres) where the effects of diachronism should be minimal. Features such as outcrop continuity of fabrics, and their relationships to major structures or to distinctive metamorphic events (migmatisation, growth of distinctively-zoned porphyroblasts) take precedence over fold style and orientation in defining individual events (cf. the change from shear zones separating undeformed areas, through asymmetric folds with strained long limbs to isoclinal symmetric folds, as inferred for $F3_k$ at Kinlochourn, section 5.2.2). The sequence of tectonic events is determined by refolding and overprinting relationships in the field and in thin section.

The numbering is descriptive, and need have no genetic significance. There need not have been a time gap, or a change in regional stress system between (say) D1 and D2 events (although this might be inferred subsequently from textural, metamorphic or geochronological evidence), and (say) D3 events of similar styles from different areas need not be contemporaneous. Because of the intensity of the later (Caledonian?) events in many areas, it is not possible to define the nature of early events with any precision - e.g D1 and D2 of a low-strain area might only be recognised as D1 of a high-strain area. Note that "highly-strained" or "highly-deformed" refers to the finite strain in the rock, not to the complexity of deformation. Highly-deformed rocks usually appear very simple, due to the elimination of planar discordances, and rotation of linear elements into parallelism with the extension direction (Escher & Watterson 1974). In some areas, pre-D1 lithological features (e.g. migmatitic banding, Moine/Lewisian interleaving) which were probably associated with tectonic structures and fabrics now only define a location fabric.

High-strain zones such as the Sgurr Beag Slide may have complex tectonic histories within them, which are not reflected outside, resulting from progressive non-coaxial deformation. Conversely, the slide fabric may become so weak away from the zone that it can no longer be recognised, giving a genuine extra phase of deformation in the high-strain zone. Thus simple matching of numbered structural sequences ($D1_a = D1_b$, or $D1_a = D2_b \therefore D2_a = D3_b$) would be meaningless.

A sketch loosely based on the Moines, showing how structural sequences could vary even without the complication of diachronism, is given in Fig. 2.9. By the end of the Precambrian, two crustal regions existed, one stabilised at 2000Ma, one at 1000Ma. The latter consisted of reworked 2000Ma basement and later metasediments. This more easterly region was the site of crustal thinning and accumulation of thick late Precambrian sediments. An early Caledonian orogeny affected this area at 500Ma, and a westerly-directed thrust or slide developed in it, cutting through the basement. As the orogeny continued, thrusts propagated westwards until at 450Ma, a major thrust placed this eastern basement on top of the 2000Ma basement and its late Precambrian cover; earlier, higher thrusts were deformed. By that time, all post-1000Ma sediments had been eroded off the area east of this western thrust. In the final stages of shortening (at 400Ma), local sets of upright folds were formed. The figure shows the sorts of structural sequences that might be recognised in various parts of this model.

In spite of such problems of correlation, useful conclusions can still be drawn from a combination of structural, metamorphic, geochronological and regional map evidence. The methods used will become apparent in Chapter 14, but essentially they involve the assumption that the regional trondhjemitic migmatisation and the main movement on the Sgurr Beag Slide can be treated as time markers. Clearly both these processes will have taken considerable geological time (tens of millions of years?), but since one of the main problems is to distinguish between Precambrian and Caledonian events, this resolution will often be adequate.

iv) Tectonic slides

As used in this thesis, a slide zone is a zone of extremely high strains associated with large displacements, having one or more slides near its centre. The use of "zone" in this context is more closely analogous to a shear zone than a thrust zone. Where a series of discrete slides occur in an area extending for several kilometres across strike (e.g. west of Ben Klibreck, Chapter 9), each having its own high strain zone and separated from its neighbour by a reasonable thickness of weakly-strained rocks, each slide has its own slide zone. In the thrust analogue (e.g. the Moine Thrust Zone) the area between the highest and lowest thrusts comprises the thrust zone. If the latter usage was applied to slides, the whole Morar Division in Sutherland would form a single slide zone (cf. Mendum 1979). Hutton (1981) falls between these two stools - he accepts the fact that a slide zone must be characterised by unusually high tectonic strains throughout, but insists on the presence of more than one break or slide within it.

The slide itself is a surface or a narrow zone separating rocks of two distinct assemblages. Where a mixed zone a few metres wide occurs, containing lenses of competent material from both sides of the slide (e.g. Morar Division psammite,quartzofeldspathic lenses from migmatitic pelite) in a schistose, micaceous matrix, it may be described as a tectonic schist. This phenomenon is relatively uncommon at the Sgurr Beag Slide, and only occurs where highly-micaceous pelites lie against it. Thus the micaceous nature of the tectonic schist reflects the nature of the materials incorporated into the melange, and not K-metasomatism in the slide zone (feldspars being altered to muscovite). Unfortunately, the term "tectonic schist" seems to carry the latter connotation for most Moine workers, and so will not be used in this thesis.

It is the fact that, on the scale of field observation, there is a discontinuity in lithology and/or metamorphic grade within the slide zone that distinguishes it from a simple shear zone. This implies either extremely high shear strains (such that kilometres of rock are attenuated beyond recognition) or brittle failure at some stage in the development of the structure. Rathbone (1980) has suggested that brittle mechanisms are required to introduce Lewisian inliers into the Sgurr Beag Slide zone. Where a slide zone has been exploited by late brittle faulting, there will also be a sharp boundary within the zone - however, it can be anticipated that the "missing" lithologies will be found in other, nearby parts of the slide zone. If the same lithologies are missing over a large area, despite wide variations in the present attitude of the slide caused by intense later folding, then any brittle faulting must have taken place early in the development of the zone, and stratigraphic separation have been enhanced by the subsequent large shear strains (themselves perhaps representing tens of kilometres movement - Chapter 7). Terms such as "ductile fault" are probably inappropriate, since although slides occur in an overall regime of ductile strain, the brittle or ductile failure of a particular rock will be affected by local factors such as mineral composition, grain size, strain rate, temperature and fluid pressure and composition.

There appear to be some differences between the Sgurr Beag Slide and Bailey's (1910) original slides in the Dalradian, which are subordinate to major recumbent folds and are associated with a much narrower zone of high deformation. They were recognised mainly from their effects on the well-defined stratigrpahy of the area, and tend to cut out only a few hundred metres of section.

Hutton's (1979a, 1979b) slides in Ireland are of this type (although he does describe some ductile features) and his type of definition (Hutton 1979c, 1981) appears to be gaining acceptance in the literature. The high strain zone at Hutton's (1979b) slide is only tens of metres wide, and strains resulting in X/Y ratios greater than 5 only occur in the inner 10-20m, with a maximum ratio of about 10 being reached in a 3m-wide platy zone adjacent to the slide. These values compare with the contact strains reported by Rathbone & Harris (1979) from the boundary between Moines and autochthonous Lewisian, or between major pelitic and psammitic units within the Morar Division, and are orders of magnitude less than those which characterise the Sgurr Beag Slide (cf. sections 7.4, 14.2). Thus most of Bailey's and Hutton's slides would not be recognised on the criteria used by Rathbone, and followed in this thesis, although a major structure like the Iltay Boundary Slide probably would be (cf. Bradbury et al. 1979, Roberts & Treagus 1979).

Slides in the Moines are more closely analogous to thrusts (especially in basement terraines) where faults are associated with zones of very high strain (mylonites) and the thrust or slide is the primary structure, any folding being secondary to it, and where large-scale inversions of stratigraphy are absent. Fleuty's (1964b) definition, which specifically links slides to large-scale folds, is thus unsuitable. Deformation in the Sgurr Beag Slide zone appears to be largely explicable in terms of simple shear, while Hutton's slide zones have much more complex strain patterns, possibly including volume changes, and the slide itself may be a means of accommodating strain incompatibilities between psammitic and pelitic units (Hutton 1981). This contrast is probably due to higher-temperature reworking of already-crystalline Moines, close to the thrust zone, and lower-temperature reworking of recently-sedimented Dalradian, further from the foreland. Some of the D1 or D2 slides mapped in the Moines by Tobisch et al. (1970) may be of "Bailey-type", having formed in the Precambrian event when the Moines were a series of sediments undergoing prograde metamorphism. These could not be recognised where, as in this thesis, only a few square kilometres were mapped in each of several widely-separated areas.

In the absence of stratigraphic criteria for slide recognition, an extension of Rathbone's (1980) criteria, derived at the Sgurr Beag Slide, will be used. These are essentially strain, presence of Lewisian inliers and contrast in lithology and metamorphic grade. This implies that any slides recognised in this thesis will be analogues of the Sgurr Beag Slide, and not necessarily of Bailey's type of slide. Highly-strained fold limbs (some of which have been called slides by O'Brien 1981 in Ardnamurchan) will simply be called high strain zones, unless they are large and intense enough to meet Rathbone's criteria.

The high ductile strain developed in this type of sliding will have eliminated any original discordances or cross-cutting relationships - thus in any areas where truncation of lithological units can be observed at the outcrop scale, the structure must be either a brittle fault, or a very minor slide with only slight displacement. Similarly, zones of moderately high strain, sufficient to eliminate cross-bedding, but not to produce platy psammites (e.g. most of the D3 and D4 zones at Mallaig) should not be called slides. About 80% bedding-normal shortening is sufficient to eliminate most cross-beds (cf. section 7.4), and such strains are not uncommon on the limbs of upright post-sliding folds (D3 of Kinlochourn, Loch Quoich, etc.).

This distinction is an important one to make, in that there is a danger of assuming, following the large displacements inferred for the Sgurr Beag Slide (section 7.4; Tanner _et al._ 1970, Rathbone 1980), that all slides in the Moines are of similar crustal importance. In fact, this is only true for the type of slide defined in this section. Thus the slides mapped by Strachan (1982) at Kinlocheil are only of minor importance, and do not rule out a stratigraphic passage from the Glenfinnan into the Loch Eil division. The zone of moderately high strain in which lies the Quoich granitic gneiss (Chapter 11) is unrelated either to tectonic emplacement of the Loch Eil Division and/or the granitic gneiss, modification of an original unconformity (cf. Lambert _et al._ 1979, Winchester _et al._ 1981) or a zone of shearing in which the gneiss was produced by metasomatism (cf. Dalziel 1966). It simply represents the normal strain state of steep $F3_q$ fold limbs, and low-strain cross-bedded psammites of the Loch Eil Division sequence can be traced through this zone, to reappear in the next major fold core to the west (Roberts & Harris in press).

The above restriction in definition is justified by the fact that this thesis deals essentially with those parts of the Moines where strain and metamorphic grade are high, and the detailed stratigraphy needed to identify Bailey-type slides has not been determined (and indeed is inapplicable if displacements in the tens of kilometres are contemplated). Nevertheless it is in precisely those areas that major structures with large displacements would be expected to occur. The point is amply demonstrated by the fact that Powell (1964, 1966, 1974), using classical criteria, identified several minor slides in the relatively low-strain area of the western Morar Division, but considered the Sgurr Beag Slide, arguably the most important structure between the Moine Thrust and the Great Glen Fault, to represent a normal stratigraphic passage at Lochailort. Only when more appropriate criteria were applied (Rathbone & Harris 1979) was it possible to demonstrate the presence of the slide, and in retrospect to find features consistent with syn-metamorphic thrusting (Powell et al. 1981).

At Kinlochourn, the Sgurr Beag Slide brings Glenfinnan Division against Lower Morar Psammite, or possibly the lower part of the Morar Schist (Chapter 5 and Tanner 1971), while at Lochailort, the Glenfinnan Division rests on the Upper Morar Psammite (Powell 1964, 1966). Johnstone (1975) and Johnson et al. (1979) inferred from this that stratigraphic separation was decreasing southwards, and that the slide was dying out (implicitly, that the Glenfinnan Division was younger than the Morar Division, and that the slide was a lag). In fact if the slide is interpreted as a thrust, and the basal Glenfinnan Division was originally unconformable on Lewisian (as suggested by the presence of Lewisian inliers in the slide zone), or at least at a deeper crustal level than the Morar Division (as suggested by the higher pressures and temperatures obtained in Chapter 15), then the younger the Morar Division rocks below the slide, the greater the stratigraphic separation across it.

Finally, now that criteria exist for the identification of high ductile strains in high-grade psammitic rocks (Rathbone 1980, Rathbone & Harris 1979) and pelitic rocks (this thesis; summarised in section 14.2), it is no longer sufficient to postulate a slide and leave unexplained the absence of both brittle and ductile features (e.g. Strachan 1982). One or the other ought to be identifiable in any structure with significant displacement. For example, if cross-bedding is preserved within a few metres of a putative slide (implying shear strains of less than 5 or 10) then this places severe constraints on the amount of ductile displacement which can have occurred. Later metamorphism might obliterate all fabric evidence for high strains, but it cannot reconstruct deformed sedimentary structures.

2.2.2 Migmatite terminology

The terminology used in this thesis essentially follows that of Mehnert (1968), modified where necessary to suit particular features of the Moine migmatites.

i) Mesoscopic features

A migmatite is considered to be a mesoscopically banded rock, in which the leucocratic portion contains significant quartz and feldspar and is of granitoid or pegmatoid aspect (granitoid meaning looking like a granite, but not necessarily having that precise composition), and in which there is reason to believe that the banding originated through igneous or metamorphic processes. The banding can be regular, irregular or lenticular (where the lenses are more than a few centimetres long), and deformed migmatites, in which the leocosome is foliated but its original texture can be deduced, are also included. Descriptive terms will generally be used, except in those chapters which deal with the origin of particular migmatites (Chapters 15, 16, 17).

The basic terminology is shown diagrammatically in Fig. 2.10(a). The leucosome is the leucocratic portion of the rock, while the remainder forms the melanosome. The leucosome is often bordered by a mafic selvedge, complementary in composition to the leucosome. The palaeosome comprises the non-migmatised portion of the rock (if any), while the neosome comprises the leucosome and (if present) the selvedge. In a closed homogeneous system, leucosome + selvedge = palaeosome, but this need not be so in practice (e.g. the palaeosome may have remained non-migmatitic because of some compositional peculiarity).

A more typical situation is shown in Fig. 2.10(b), based on sample Q22 (cf. section 16.3.2). The coarse vein contains biotite-rich schlieren (possibly fragments of selvedge) while the host rock contains lenticles of leucosome, on a scale too fine to be described as migmatisation. However, it is still useful to use the terms leucosome (closest to the melt/segregation composition) and palaeosome (closest to the original bulk composition). In this particular case of partial melting, the new leucocratic material is simply described as "melt", and the refractory residue as "restite". In subsolidus migmatites, the leucocratic portion is described as the "segregation". Mobilisation carries no implication of melting or segregation, and may also be applied to near-complete melting or recrystallisation, where all structures break down ("highly-mobilised").

ii) Megascopic features

The most common type of migmatite in the Moines shows sub-planar, sub-parallel banding of leucosome and melanosome (Fig. 2.11(a)). This is the stromatic (layered) type of Mehnert (1968), but a term used more commonly in the Moines is lit-par-lit (cf. the lit-par-lit injection of Michel-Levy 1893). Note that the use of the term in this thesis does not imply injection (since the Moine migmatites traditionally described thus were produced by local subsolidus segregation - section 17.1). Lit has been used in Moine literature to denote a leucocratic vein in a lit-par-lit migmatite (e.g. Johnstone 1975, p.40).

Agmatitic (breccia) structure is analogous to that of an igneous breccia (Fig. 2.11(b)), angular blocks of melanosome being veined by leucosome. An agmatite is an agmatitic migmatite, and the leucosome is said to have agmatised the melanosome.

Schollen (raft) structure is similar (Fig. 2.11(c)), but the proportion of leucosome is greater, and the fragments of melanosome are usually elongate parallel to foliation or banding.

Diktyonitic (net-like) structure (Fig. 2.11(d)) is produced when mobilisation takes place in shear zones (e.g. in the granitic gneiss, section 17.3.2); however, the conjugate sets of shear zones necessary to produce a genuine net-like structure are rarely found in the Moines.

Mehnert's (1968) phlebitic (vein) and folded structures simply describe deformed lit-par-lit migmatites (Fig. 2.11(e),(f)).

Augen (eye) structure occurs where coarse-grained feldspars are wrapped by finer-grained micas and/or quartz, generally as a result of deformation (Fig. 2.11(g)). It is also possible to describe individual lits as being augened.

Schlieren structure describes the situation where the rock consists entirely of leucosome and biotitic wisps or streaks - usually as an extreme development of lit-par-lit texture (Fig. 2.11(h)). Mehnert's use of the term carries an implication of laminar flow in the material, but an identical structure could be produced by the coalescence of leucosomes and selvedges in a previously-foliated rock.

Nebulitic structure describes a highly-mobilised rock in which the boundaries of the melanosome are more diffuse, and the melanosome itself is quite leucocratic.

Finally, ptygmatic veins are discrete continuous veins which display tight buckle folding, often with negative interlimb angles, indicating that the vein had much greater strength than its country rock during shortening along its length.

2.2.3 Stratigraphic terminology

In general, lithostratigraphic terms have been used, essentially following I.G.S. practice. The largest units or divisions are in part tectonic, the Morar Division being separated from the Glenfinnan Division by the Sgurr Beag Slide (Tanner et al. 1970), but they also have distinct lithological characteristics (Johnstone et al. 1969). Johnstone (1975) recommended restricting the term formation (e.g. Upper Morar Psammite Formation) to western Inverness-shire where tectonic strain is relatively low. Some of those units may yet prove to be tectonically bounded (e.g. the Knoydart Pelite of Chapter 4, which was previously considered to lie in a stratigraphic succession - Ramsay & Spring 1962), so in general units will be referred to simply as Upper Psammite, etc. Informal names, e.g. the Lochan Breac Pelite and Lochan na Bracha Psammite of Chapter 6, are of only local application, although those at Acharacle correspond to the units in O'Brien (1981). In some small areas, units have simply been designated "migmatitic pelite", "non-migmatitic psammite", etc.

Terms such as "pelite", "psammite", etc. are for field use and so not precisely defined; they essentially follow I.G.S. usage, as inferred from published maps. In practice, psammites have less than about 5% mica+garnet, micaceous psammites 5-10%, semipelites 10-30% and pelites >30%. It is important to note that many rocks described as pelites in this system are not metamorphosed shales, and that "pelite" covers a broader compositional range than in many other metamorphic terraines. The typical migmatitic pelite of the Glenfinnan Division and the garnetiferous pelite of the Morar Division have greywacke or siltstone compositions (cf. Butler 1965 and sections 16.1, 16.2 of this thesis) - this probably reflects the old interpretation of the quartzofeldspathic material as allochthonous, where the micaceous layers (restite on an isochemical model for migmatisation) represented the original sedimentary composition. The term "pure pelite" is sometimes used to distinguish a rock with a very aluminous (micaceous) bulk composition.

2.2.4 Units, map references and sample numbers

i) Field

Map references are from the U.K. Ordnance Survey National Grid and are specified by two letters (defining the 100km square) and a six- or eight-figure number (defined to 100m and 10m respectively), the first half being the easting and the second the northing; they are enclosed in square brackets in the text - e.g. [NM 563271]. Within a given small area, the letters may be omitted after the first few references.

All azimuths are measured relative to Grid North, and are always written as a three figure number (N=000, E=090, S=180, W=270). Dips of planes and plunges of lines are measured from the horizontal (see also section 2.2.1(i)).

Distances and dimensions are given in S.I. units: metres, km, cm, mm, dm (decimetre = 0.1m) etc.

ii) Laboratory

Dimensions are given in metres, etc. or microrns ($1\mu = 10^{-6}$m), weight in grammes, kg etc. Analytical results are given in weight percent for major elements and parts per million (ppm) for traces. Rare earths were normalised by dividing the analysed values by those from the Leedey chondrite (Masuda et al. 1973). In some of the thermodynamic calculations, S.I. units have not been used, in accordance with normal geological practice. Temperature is presented in °C (0°C = 273.15K) and pressure in kilobars (1kb = 10^8Pascals). In some cases, calories have been used for energy, rather than joules (1cal = 4.184J).

All samples have been given local numbers reflecting the area of collection (e.g. BK3 from Ben Klibreck). All samples from which thin sections were taken have been given Liverpool University numbers (e.g. 54323), which will always be quoted as 5-figure numbers in round brackets. In chapters dedicated to a particular area, only the Liverpool number will be quoted; otherwise, the sample will be referred to by its local number to emphasise the area from which it was collected. In Appendix 4, all samples which were referred to are listed by area, with their Liverpool numbers and a grid reference for each locality.

2.3 Analytical techniques

Whole rock analyses were carried out using standard powder pellet techniques at Liverpool University (1979/80). Samples were cleaned by grinding and reduced in two stages to <1mm using a steel jaw crusher. The fragments were then crushed in either a steel tema (for a total of 2-3 minutes) or an agate ball mill (for several hours). Sub-samples of about 5g for use in FeO determinations were removed from the tema after 40 seconds (to avoid excessive oxidation) and crushed by hand in an agate mortar to 150 microns. Before putting samples into the ball mill, a split was removed and crushed by tema for FeO determinations. On removal from the ball mill or tema, 7g of each sample was ground by hand to 53 microns, and used to prepare powder pellets (pressed under 4-6 tons, with moviol as a binding agent). At each stage, the powder was thoroughly homogenised, and the unused splits are stored at Liverpool University.

These pellets were then analysed by XRF, along with appropriate standards, and the program of Brown et al. (1973) used to reduce the data to weight percent of major elements and parts per million of traces (total Fe being quoted as Fe_2O_3). Only those samples which had been crushed in the ball mill were analysed for Co, Cr and Ni (cf. Appendix 3). To test for contamination, the granitic gneiss Q11a was crushed in both the ball mill and the tema (FeO being determined on the agate-crushed powder, rather than a separate split). It is apparent from Appendix 3 that only Cr was significantly increased by the tema, with Ni perhaps showing some slight contamination; the differences in SiO_2 and Fe_2O_3 are probably within errors, but the agate-crushed sample is severely oxidised.

FeO was determined using standard wet methods. 5g samples were ground to 150 microns and dissolved in hot, concentrated HF + H_2SO_4 solution. This was added to a mixture of sulphuric, boric and phosphoric acids (to maintain acidity during the titration, to complex F^- and to remove the effect of Fe^{3+}) and titrated with $K_2Cr_2O_7$ solution, using sodium diphenylamine sulphonate indicator.

Duplicate analyses were performed, and only accepted if they came within ±0.02 wt. %. FeO was converted to Fe_2O_3, then subtracted from Fe_2O_3 in the tables, with the overall total adjusted accordingly. CO_2 and H_2O were not determined. Microprobe analyses were carried out on polished sections prepared at Liverpool University, using Manchester University's Cambridge Geoscan instrument fitted with a Link Systems energy-dispersive spectrometer. The data were reduced on-line by an iterative peak-stripping program incorporating ZAF corrections. These are standard Manchester University methods, and are an updated (1980) version of the procedure given in Dunham & Wilkinson (1978). These authors showed that precision and accuracy compared favourably with wavelength-dispersive methods for elements with concentration greater than 1 wt. %, and that a large part of the analytical error (65-75%) was due to the counting statistics. The 2σ counting error is quoted on the computer output, and has been used to give minimum errors in Chapter 15.

Rare earth determinations were carried out by M.S. Brotherton on "FeO" powders, using neutron activation at the Risley reactor centre and the standard Liverpool University program to calculate the ppm of REE.

CHAPTER 3

Mallaig

3.1 Introduction

Mallaig lies on the northwest limb of the Morar Antiform, 10km southeast of the Moine Thrust, and 30km west of the nearest trace of the Sgurr Beag Slide (Fig. 1.1). The area was chosen as a representative of the non-migmatitic Morar Division of the southern Moine outcrop. Its structure and metamorphism were studied in detail to develop a model which would be generally applicable west of the Sgurr Beag Slide.

It is the type section for the Lower Psammitic Group and the Striped and Pelitic Group of Richey & Kennedy (1939), and also includes the lower part of their Upper Psammitic Group. The rocks lie on the vertical-to-overturned western limb of the Morar Antiform, but are often only weakly deformed, preserving numerous sedimentary structures (Fig. 3.1).

3.2 Stratigraphy

Three major lithostratigraphic units are present: the Lower Morar Psammite, the Morar (Striped and Pelitic) Schist and the Upper Morar Psammite (Figs. 3.1, 3.2 - terminology of Johnstone 1975). Abundant cross-bedding indicates consistent westward younging, in agreement with Richey & Kennedy (1939). Although this sequence is disrupted by several zones of high D3 and D4 strain, only that within Unit 4 is considered large enough to have significantly disrupted the stratigraphy.

Subdivisions of the Morar Schist will be referred to informally as the Upper Striped Schist, Garnetiferous Pelite and Lower Striped Schist. For descriptive purposes, they have been further broken down into numbered units of only local significance. The sequence is summarised below, and lithological details expanded in Fig. 3.2.

3.2.1 Mallaig

Lower Morar Psammite

This unit comprises thick (1-2m) feldspathic psammite beds, separated by thin (10-20cm) semipelites. Undeformed planar and trough cross-bedding is present in most beds, indicating consistent westward younging. The base of some beds is rich in ½-1cm quartz and feldspar pebbles. This corresponds in position and lithology to Richey & Kennedy's (1939) Unit (b); only the upper 150m was mapped.

Morar (Striped and Pelitic) Schist

a) Lower Striped Schist

Unit 1 This is a transitional lithology, showing an upward decrease in thickness of psammite beds from 1m to about 30 cm, with semipelites increasing to a maximum of 50cm. This unit is highly deformed (no cross-beds, strong foliation parallel to bedding).

Units 2-4 These are dominantly muscovite-rich pelites with rare garnet; 50cm thick cross-bedded psammites with pebbly bases become common in the upper part. Feldspar pebbles are up to ½cm in diameter and subrounded; one bed contains 5-10cm pebbles of micaceous psammite at its base. Unit 4 is very highly deformed, and may mark the site of a minor tectonic slide.

Units 5-7 Units 5 and 6 consist of finely striped (a few centimetres) pelitic and semipelitic schists. Pelites are muscovite-rich, semipelites biotite-rich. Towards the top of Unit 6, psammite bands a few decimetres thick appear, while Unit 7 is a thick, highly-deformed micaceous psammite.

b) Garnetiferous Pelite

Unit 8 This comprises a thick, homogeneous pelite, which has abundant large garnets (>1cm). The schists are lenticular on the scale of a thin section, the lenses consisting of: muscovite with abundant garnet and lesser biotite; semipelite with quartz, biotite, plagioclase and small garnets; coarse quartz-plagioclase segregations. In the west, thicker semipelitic bands appear, and the rocks grade into Unit 9.

c) Upper Striped Schist

Unit 9 consists of homogeneous semipelite with small garnets. Calc-silicates appear at this horizon, as bands a few centimetres thick and up to 100m long. They are strongly symmetrical, with a 3cm calc-silicate flanked by 6cm micaceous psammite ribs. In detail, mineralogy varies gradationally about the centre, with calcite, apatite and zoisite decreasing, while muscovite and clinozoisite appear. Their thinness, symmetry and persistence, allied to the concentration of CaO, CO_2 and P_2O_5 in the core, make a diagenetic (concretionary) origin likely.

Units 10-12 Unit 10 is a thick, flaggy, micaceous psammite. It grades upwards into a striped psammite/semipelite assemblage, then into Unit 12, a poorly exposed series of striped semipelites. Calc-silicates are common throughout.

3.2.2 Glasnacardoch Bay

Two sections are seen here, with slight differences in their stratigraphy. Units 14 to 16 are traceable continuously at low water. Unit 12 is similar to the rocks along strike in East Bay, Mallaig, and the base of the Upper Psammite is everywhere well-defined, although it lies in a high strain zone in the southern section.

Northern section

Units 13-15 are dominantly semipelitic, but include a thick, muscovite-bearing psammite showing high D3 and D4 deformation.

Units 16-17 consist of banded (10cm) psammite and semipelite. Calc-silicates are abundant in Unit 16, but become less common higher up the sequence.

Units 18-21 Unit 19 consists of psammite, biotite semipelite and muscovite pelite interbanded on all scales; some psammites are cross-bedded, and calc-silicates are common. Unit 20 is a homogeneous pelite, carrying rare calc-silicates, while Units 18 and 21 are thick, highly-deformed psammites.

Unit 22 This is a thick sequence of laminated semipelites, calc-silicates and 10cm-thick psammites, these last showing small-scale cross-bedding, ripple-lamination and soft-sediment deformation. Mineralogy is more variable than elsewhere, and microcline porphyroblasts are common. Typical biotite semipelites and muscovite pelites contain up to 10% microcline, but a distinctive semipelitic rock, rich in microcline and muscovite, with lesser quartz, biotite and garnet, also occurs (e.g. 54323).

Southern section

Units 13a-c Unit 13 can be divided into a lower, undeformed psammite, a homogeneous semipelitic schist, and a muscovite-rich pelitic schist.

Units 14-16 are identical to those in the northern section.

Units 17a-19a Unit 17 can be divided into a thick, homogeneous pelite, and a banded psammite/semipelite. It is followed by a highly-deformed psammite, then a banded psammite and semipelite, with abundant calc-silicates.

Units 21a-22a These consist of a thick psammite, followed by finely banded psammites, pelites and semipelites (often microcline-rich); all are highly-deformed.

Upper Morar Psammite

This group is marked by the incoming to thick (>1m), massive, cross-bedded psammites, and restriction of semipelites to thin (<20cm) intercalations. The psammites are very variable lithologically, may be quartzitic, feldspathic, muscovite- or biotite-rich, and contain heavy-mineral bands or lenses (magnetite-biotite-sphene-zoisite-zircon-muscovite-quartz). Semipelites are biotite-rich, and may contain either microcline or quartz as the dominant light mineral. The two types are often finely interbanded (e.g. 54322). In the southern section, the lower 20m is very highly strained, but in the northern section the rocks are virtually undeformed, and a clear sedimentary transition is seen.

3.2.3 Correlation

From the map (Fig. 3.1), it is likely that the top of the Mallaig section lies close to the base of the Glasnacardoch section; this is supported by the presence of isolated exposures similar to Unit 12 along the roadside. Units 13a and 17a are similar to 13 and 17 overall, but in the southern section, the lithologies are less intimately mixed. The distinctive microcline-rich rocks confirm correlation of Units 22 and 22a, so it is only in Units 18 to 21 that any doubt exists.

3.2.4 Igneous rocks

Two sets of dykes were emplaced sub-parallel to bedding - late Caledonian appinites (Dearnley 1967, Smith 1979), and dolerites, presumably Tertiary.

Appinites These form massive, homogeneous dykes, post-dating all ductile deformation. They are coarser than the dolerites, show some marginal chilling (but still with 1mm grain size), and have a blocky weathering pattern, lacking the dolerite cooling joints. In thin section, 2mm euhedral zoned hornblendes and rarer biotites are poikilitically enclosed by normally-zoned, tabular plagioclase crystals. The latter pass out into (?replacement) patch perthites, with zoned plagioclase relics in alkali feldspar. This passes out into homogeneous, untwinned alkali feldspar, which is continuous with the interstitial alkali feldspar/quartz mesostasis. A few chlorite patches occur, but the rocks seem essentially unmetamorphosed.

Dolerites These are typical ophitic dolerites, with very fine-grained chilled margins; they cut all structures, and are often emplaced along N-S faults as well as parallel to bedding.

3.3 Structure

3.3.1 Soft-sediment deformation

This is most marked in the Upper Morar Psammite where tight-to-isoclinal folds of bedding-lamination occur within beds, with no effect on upper or lower bed surfaces and no inversion of cross-beds. Most folds have an approximately similar style (Class 2), with thinned limbs, and some show refolding or truncation by later cross-bedding. Oversteepening of cross-bedding in the same unit is probably also sedimentary in origin; all these features point to a northwards and upwards transport direction, in agreement with other palaeocurrent indicators.

Elsewhere, refolded isoclinal folds in laminated psammite/semipelite of Unit 22 are probably sedimentary in origin, since only a weak S1 schistosity is present and fine-scale ripple-lamination is preserved. Throughout the Morar Schist, small-scale tight-to-isoclinal folds, transected by the S2 fabric, could be of sedimentary or F1 (tectonic) origin.

3.3.2 Tectonic deformation

Six main phases of deformation were recognised. The first two gave rise to a penetrative schistosity, and tight-to-isoclinal folds. D3 and D4 structures are restricted to localised zones of high strain, sub-parallel to bedding. These fabrics are deformed by open, gently-plunging folds with flat-lying axial planes (F5), and upright E-W folds and kink bands.

(i) D1

The widespread bedding-parallel S1 fabric (Plate 3.1(a)) is most clearly seen in micaceous psammites, where it is defined by shape and crystallographic alignment of dispersed 0.2mm micas. In pelites it is a penetrative muscovite/biotite schistosity (recrystallised in MP2), carrying 0.5x0.05mm plates of ilmenite (similar to those included in garnets). Several mesoscopic F1 folds occur in the Lower Striped Schist, being defined by 1m psammites separated by 2-3m of pelite. These are reclined isoclines, with an essentially parallel style in psammites, and have a *fanning pressure solution cleavage. The small folds* mentioned in section 3.3.1 may be F1 or sedimentary.

F1/F2 relations can be seen in the interference pattern of Fig. 3.3(a) where a penetrative S1 schistosity, oblique to bedding in an F1 fold core, is overprinted by F2 crenulations at almost 90° (although elsewhere, this angle is usually smaller). No large-scale F1 folds were recorded, and the consistent younging and absence of stratigraphic repetitions argues against their presence in this area. F1 apparently faced eastwards prior to F2 folding, in agreement with the conclusions of Powell (1974).

(ii) D2

S2 is the dominant schistosity in this area; it is usually penetrative, but in thin pelitic laminae between psammites, is often a crenulation cleavage, and intrafolial folds are common (Fig. 3.3(c)). Its progressive development from S1 can be seen in pelite/semipelite transitions (Plate 3.1(a)), where F2 crenulations nucleate on MP1 garnets in the pelite. The S2 fabric often has associated a steeply N-plunging lineation - it is defined by an LS shape fabric in quartzites, and in pelites by rodded quartz veins, mica elongation and long axes of quartz segregations. F2 folds are common on all scales, from a thin section to 10m wavelength, and range from close to tight, with axial planar S2 cleavage (Plate 3.2(c)). On a large scale, tight folds with a few metres wavelength form Type 3 interference patterns with F1 and F3 folds in Unit 3 (e.g. Fig. 3.3(a)), and a Z-profile fold of about 10m wavelength occurs in Unit 5 (at [685978]). Over most of the area, S2 strikes clockwise of S-S1 and dips steeply east, while bedding dips steeply west (Fig. 3.4(a)); the rocks face northwards and upwards in S2. The one exception is Unit 3, where F2 folds have S-profiles. These rocks face southwards in S2, indicating that a simple F2 fold could not be responsible. Stereographic analysis suggests that the change could be due to a rigid rotation about a horizontal, E-W axis (perpendicular to bedding), possibly related to an overlying slide (Fig. 3.5(a)). Alternatively, F2 folds may have been conjugate locally.

(iii) D3

This is usually represented by a strain-slip schistosity, sub-parallel to bedding in pelites and semipelites; folds were only recorded in Unit 3, where they range from close to isoclinal. The S3 schistosity varies markedly in intensity across strike, zones in which it is penetrative alternating with others in which it is virtually absent. On Fig. 3.1, an indication has been given of the intensity of S3 and S4. S3 is generally restricted to pelites and semipelites, while S4 may be present in any rocks. They have not been distinguished on the map, as positive identification can only be made in thin section. Outside Unit 3, S3 has the form shown in Plate 3.1(b). It intersects S2 at a low angle, but S2 shows little reorientation within microlithons, so that S3 cleavages resemble miniature shear zones (cf. S2, which develops via a crenulation, so that S1 is reorientated). Within Unit 3, S3 is generally less penetrative, and in the cores of F3 folds is a typical crenulation cleavage. F3 crenulations tend to be tight and angular, with partial recrystallisation of micas; their age is confirmed by their relations to S2 striping and to outcrop-scale interference patterns. Large F3 folds are tight-to-isoclinal, tending towards Class 1B in psammites, Class 2 in pelites (Plate 3.2(b), (a)). Axes plunge steeply to the south when bedding is folded, more shallowly when S2 is folded - presumably reflecting initial differences in orientation (Fig. 3.4(a)).

(iv) D4

This phase is only represented by an S4 fabric, which takes the form of grain-size reduction zones, sub-parallel to bedding. Like S3, its intensity varies across strike; it is the dominant fabric in certain highly-deformed psammites (cf. Plate 3.3(a), (b)). The textural relations of S4 are best seen in semipelitic schists carrying S3. In (54094), Plate 3.1(b), the zones clearly cut and displace the S3 schistosity, and in this case are about 0.5mm wide; they correspond to a weak, bedding-parallel cleavage in hand specimen. The nature of the zones can be seen in (54092), Fig. 3.3(d). Quartz in the centre of the zone forms 0.01-0.02mm grains, or 0.05mm grains with 0.01mm subgrains; this is flanked by a 0.5mm wide zone in which matrix quartz ($>$0.1mm) has sutured grain boundaries and smaller subgrains, and by a more diffuse zone in which matrix quartz shows undulose extinction. Aligned pale green chlorite occurs in the centre of some zones, and along with irregular plates of haematite probably arises from breakdown of matrix biotite. (54107) is a fine-grained psammite ($<$0.05mm), containing numerous smaller subgrains of quartz, with ragged muscovites similar to those in mylonites, and tiny new muscovites and chlorites. It comes from one of the highly-deformed psammite beds, and probably represents coalescence of these zones. On a mesoscopic scale, S4 fabrics are folded by F5 and F6, confirming their position in the sequence. A possible mesoscopic S4 zone was seen in the gully at [68479773], Fig. 3.3(c). Here, what seems to be the S3 fabric (oblique to bedding) is deformed by a bedding-parallel shear zone, and a new fabric developed. Due to deep weathering, thin sections could not be obtained to confirm indentification of the fabrics; the apparent displacement indicates a sense of movement lying between dextral, and eastern side downthrown. The microstructure of S4 is certainly consistent with their being mini-shear zones, the fine-grained centres corresponding to the highest strain (or strain rate).

(v) D5

This phase is represented by open minor folds deforming S1-S4 (Plate 3.3(c), (d)), and associated crenulations. The folds are most common around Glasnacardoch Bay, although they are present in the north. Axial planes dip about 20-30° to the east or northeast; axes plunge in the same direction (Fig. 3.4). Fold style ranges from concentric to chevron. The folds are strinkingly cylindrical, forming mullions which can be traced for tens of metres, but are disharmonic, often with curviplanar axial surfaces. Quartz (±chlorite±calcite) veins are common, both dilational veins or pods on fold crests, and tabular veins parallel to axial planes. F5 crenulations are open, and bend or break earlier micas (Plate 3.1(c)). They locally tighten, and develop a strain-slip cleavage, along which minor new white mica grew.

(vi) D6

This includes a wide range of features - open E-W folds and crenulations, and kink bands of several orientations. The latter dominate in the Upper Striped Schist, probably due to its laminated nature (Plate 3.2(d)). Medium-scale F6 folds are open, with E-W axial planes and steeply WSW-plunging axes. They are associated with open-to-close crenulations of S2 (Plate 3.1(e)). These are often concentric, and die out along their axial traces. Micas are bent or broken, and some pressure solution of quartz and calcite is indicated by their concentration in crenulation cores. The major structure in Glasnacardoch Bay has the form of a large kink band, although the southern boundary is sharper than the northern. This area provides the best evidence for the relative ages of F5 and F6. S-S1, F5 axes, and F5 axial planes were plotted for the subareas shown on Fig. 3.1. It is clear from Fig. 3.5(b) that F5 axes and axial planes have been rotated with bedding, indicating that they predate the map-scale F6 structure.

3.3.3 Discussion

The orientation of structural elements in the northern and southern areas has been plotted on Fig. 3.4. Relevant features were pointed out in the above sections. The number and distribution of readings do not justify contouring, and S-S1 and S3 have not, in general, been distinguished, as they are assentially parallel at most localities.

The most obvious feature is the geometrical relationship of S-S1 and S2, which would give a steep ENE intersection, similar to the observed lineations; the southward decrease in L2 plunge, and the contrast in plunge of F3 folds of bedding and of S2, are also clear. In comparing Figs. 3.4(a) and 3.4(b), the southward shallowing of L2 is parallelled by the decreasing strike divergence of S-S1 and S2. The spread of S-S1 and S3 points in Fig. 3.4(b) is due to the major F6 structure in the bay, and confirms its near-vertical axis.

F5 axes in general plunge at about 20-30° to the north or northeast. In a narrow zone, parallel to a 140-trending cliff, F5 axes and axial planes steepen in about 20m to 60°, remain steep for 70m (subarea V, Fig. 3.5(b)) then return to normal. Sericitisation is intense in this zone. Movement of both elements seems to be in a vertical, NE-SW plane - it could be due to a flexure on a 140-trending axis, or a steep, 140-trending shear zone, with dip-slip movement. Assuming that a late, open flexure would have a symmetrical axial plane (dipping moderately SW), the vertical boundaries to the zone suggest the latter explanation. It is not possible to determine the timing of this structure relative to F6.

Both D3 and D4 give rise to bedding-parallel zones of strong ductile deformation; both microscopically and macroscopically, S3 and S4 resemble shear zones rather than conventional axial planar foliation. Major slides (e.g. the Knoydart and Sgurr Beag slides) have identical fabrics associated with them, but more intense, and covering a wider area (e.g. psammites here are platy on a scale of 5mm rather than 1mm - cf. sections 4.2, 7.4). The narrow zones of platy psammite (which appear subjectively as deformed as the outer parts of major slide zones) probably do not seriously affect the stratigraphy. There is no way of estimating the strain in the platy pelites of Unit 4; pelites become platy before cross-beds are lost, and further increases in strain are not recorded (cf. section 7.4). The localisation of F3 folds below this zone (section 3.5) may indicate the presence of a significant slide, but the zone is still less than 1/10 as thick as those associated with major slides.

3.4 Metamorphism

Textural evidence for timing of crystallisation of the major phases will be described first, then a metamorphic history suggested.

3.4.1 Matrix minerals

Quartz, feldspar The main crystallisation of quartz was MP2 in age. It gave rise to the 0.1-0.5mm annealed texture in rocks carrying S1, F2 crenulations or S2. Pre-MP2 quartz may be preserved as inclusions in MP1 garnets (section 3.4.2) and as coarser grains in veins (section 3.4.3). This annealed quartz is deformed and recrystallised in rocks showing F3 folding or carrying S3 or S4 fabrics (Plate 3.1(b)). In these cases, quartz grain boundaries become sutured, undulose extinction and subgrains develop, and finally new smaller grains appear - presumably representing essentially strain-induced recrystallisation. D5 has little effect on quartz, possibly causing some undulose extinction, while concentration of quartz in F6 cores indicates significant pressure solution. Matrix feldspar (plagioclase about An_{20}) shows the same relations as quartz to D1, D2 and D3; its relations to D4 are not clear, but it is sericitised where involved in D5 or D6 deformation.

Micas Biotite and muscovite are usually intergrown to form the schistosity. Again, the main crystallisation is MP2. Where preserved, S1 is defined by coarse biotite and muscovite, but these could have grown mimetically after early chlorite. Micas in F2 crenulations are polygonised and unstrained (Plate 3.1(a)). Micas in D3 crenulations are only recrystallised where the latter are tight, but new micas have grown in the S3 schistosity (Plate

3.1(b)). Only the most strained (highest energy) micas have recrystallised, suggesting that temperatures were lower than during MP2, when universal recrystallisation occurred. Limited fluid access during D3 could have a similar effect, and might be expected if D3 represents the onset of Caledonian reworking of an earlier metamorphic complex (section 14.4). S4 zones cut micas with no apparent recrystallisation, and micas are broken or bent by F5 and F6 crenulations (Plate 3.1(b), (c), (e)).

Chlorite Matrix chlorite usually replaces biotite or muscovite. It is often spatially associated with S4 zones, with MS5 veining, or with fractures or veins related to F6.

Calcite This is always late, and is present as large, unstrained crystals extending along matrix grain boundaries, or concentrated in cores of late folds.

Zoisite/clinozoisite These minerals occur in a few semipelites, and in calc-silicates. They form tiny (0.01-0.02mm) rounded prisms, are included in matrix minerals, and must have formed at least early in MS2. Clinozoisite is common in MP1 segregations, possibly indicating a similar age for matrix examples. In semipelites, clinozoisite with about 5% total iron (as FeO) is present, but calc-silicates contain zoisite with about 1% total iron (Appendix 2, sample MAL51).

Ores The dominant opaque mineral is ilmenite, occuring both as plates lying in S1 and deformed by F2 crenulations, and as widely distributed granules. Haematite is common in rocks which have sufferred retrogression.

3.4.2 Porphyroblasts

Garnet This is the most common porphyroblastic mineral, being present in all rock-types, and reaching 1cm diameter in the Garnetiferous Pelite; all are essentially calcic almandines (Appendix 2). Garnet growth is dominantly MP1 in age. Garnets overgrow the S1 schistosity in semipelites (Plate 3.1(a)), with continuous inclusion trails of quartz and ilmenite. In psammites, garnets are skeletal and include matrix-sized quartz grains, suggesting some MP2 growth. Zoning is common (cf. MacQueen & Powell 1977), with the idioblastic zone boundary marked by a sharp increase in the size and density of inclusions, which then decrease towards the margin (Plate 3.1(d)). Often, the outer zone has a curved inclusion trail, continuous with a straight trail in the central zone, indicating an element of MS2 growth. As a strong S2 schistosity develops, continuity of S_i with S_e is lost, due to rotation and recrystallisation. In the Garnetiferous Pelite, two sizes of inclusion-free garnet are present in the muscovite-rich lenses. The large garnets clearly predate at least part of the D2 deformation; the smaller ones are of similar size to the matrix micas, hence conventional criteria for age determination cannot be applied. However, they display similar chemical zoning to the large garnets, so are probably of essentially the same age (Appendix 2). Garnets are overgrown or embayed by MP2 biotite and microcline, and are strongly augened by S3. They are retrogressed to chlorite adjacent to S4 zones, and in Plate 3.1(c) a clear example is seen of a garnet with curved S_i incongruent on F5 crenulations.

<u>Microcline</u> These porphyroblasts are common in semipelites near the top of the Upper Striped Schist and base of the Upper Morar Psammite. They tend to occur in muscovite-rich layers, but not exclusively so. The porphyroblasts are xenoblastic, equant, ½-1mm in diameter, with abundant tiny droplike inclusions of quartz, rare plagioclase and idioblastic biotite. They may be perthitic (0.1mm rods), and are rarely twinned, but optically all are microclines. They cut S2 and overgrow F2 crenulations. The first structures seen to deform them are F5 crenulations (Plate 3.1(c)), but the most likely age is MP2, synchronous with the main recrystallisation of the other minerals (as was suggested by Smith & Harris (1972) for similar porphyroblasts in South Morar).

<u>Biotite</u> Biotite porphyroblasts are rare but widespread. Some cut the S2 crenulation cleavage and overgrow MP1 garnets, but others are augened by S2; the later ones are deformed adjacent to an S4 zone. A few occur parallel to F3 crenulation axial planes.

Muscovite MP2 muscovite porphyroblasts occur with biotite, cutting F2 crenulations and parallel to S2. Muscovites also cut all later crenulations. Their restriction to hinges suggests a causal relationship - e.g. they may have nucleated in highly strained or fractured micas. There were probably minor episodes of porphyroblastesis after each deformation phase.

Chlorite Chlorite occurs as sheaves cutting the S2 or S3 foliations. In some cases their origin can be related to a nearby S4 zone, or a vein out of which they grow into the matrix. Two species of chlorite occur. One, common in the Garnetiferous Pelite, is very pale yellow, with α absorption $>\gamma$, is length fast with small positive 2V, and shows normal grey interference colours. It has a fairly high refractive index (>1.6). The other has low relief and is identical to the common vein chlorite; it is strongly pleochroic (α pale yellow-green, γ lime green) and is length slow, with a small negative 2V. This chlorite also occurs as an alteration product of garnet, and as spherulitic mats which locally develop into good crystals. The two types are sometimes intergrown, but in (54096) the length-fast chlorite is kinked while the other is not, suggesting that the former may have a shorter time range (e.g. only MS4).

3.4.3 Segregations/veins

Quartz veins These were emplaced throughout the deformation history. The main sets are:-

a) parallel to S-S1, folded by F2

b) post-S2, pre-F3

c) syn-F3 - many veins in zones of strong S3 are probably of this age, as are veins and pods disrupting psammites in F3 fold cores in Unit 3

d) post-S3, oblique to bedding, cut by vein parallel to F6 fractures

e) parallel to F5 axial planes, and on F5 fold crests (Plate 3.3(c))

f) parallel to fractures related to F6, and to F6 axial planes.

Quartz-calcite(-chlorite) veins A quartz-calcite vein cuts S2 in (54082), and quartz-calcite-chlorite veins form on F5 fold crests. Various combinations of quartz, calcite and chlorite fill bedding-parallel fractures, which are common in zones of strong kinking and probably related to F6. Carbonate-bearing veins in pelites seem to indicate lower grade than plagioclase-bearing veins (Vidale 1974), so their restriction to post-D2 may be significant.

Quartz-oligoclase veins These are uniformly post-D1 and pre-D2, and so are restricted to the metamorphic peak. They are dominated by quartz, but contain plagioclase (An_{20}), with minor muscovite, biotite and garnet. They are frequently folded by F2, and are sometimes oblique to S1. Thick veins usually show little recrystallisation, but thinner ones are reduced to near matrix grain size; the new plagioclase is also An_{20}.

Quartz(feldspar-mica-clinozoisite) segregations Two groups were recorded:-

a) Small (1mm) segregations in Unit 22 semipelites. These are sub-spherical, have coarser quartz than the matrix (0.5mm), and contain unorientated biotite, minor microcline and included 0.01mm clinozoisite prisms. Oligoclase and muscovite are sometimes present. In pelite, the S1 fabric is compressed around the segregations to a similar degree as with MP1 garnets, and the segregations are more elongate, with biotite parallel to the foliation, and quartz recrystallised to 0.15mm. However, the wide variation in elongation (up to 7:1), in rocks whose garnet/S1 relations indicate about 50% shortening, shows that some at least were originally lenticular. They predate at least part of the MP1 garnet growth, and probably formed early in MP1 - they are analogous to Smith & Harris's (1972) pre-D2 microcline "porphyroblasts".

b) Small (few millimetres) quartz-rich lenses, parallel to S2 in pelites of Units 5 and 8. These segregations are wrapped by strong S2, yet contain large, unstrained crystals of quartz, with minor oligoclase, microcline or mica. They are strongly elongated parallel to F2 axes, and are probably broadly MS2.

3.4.4 Metamorphic history

Metamorphic grade during D1 has been obscured by later recrystallisation. Although the S1 foliation now contains biotite, this could have grown mimetically in MP1. Only quartz and ilmenite inclusions in garnet are relics from MS1; their fine grain size could be taken to indicate relatively low grade.

MP1 metamorphism involved the growth of garnet porphyroblasts, quartz-oligoclase veins containing minor mica and garnet, and quartz-oligoclase-microcline-biotite (-muscovite-clinozoisite) segregations.

MS2 grade is indicated by syntectonic garnet growth, and development of a muscovite-biotite schistosity; the increase in inclusion size at the garnet zone boundary presumably reflects the inclusion of MP1 as opposed to MS1 quartz.

MP2 metamorphism caused the main matrix recrystallisation (of quartz, oligoclase, microcline, biotite, muscovite), and the growth of biotite, muscovite and microcline porphyroblasts. Some skeletal garnet may have grown, and the narrow Mg-poor (?lower temperature) rims of garnets may also be of this age (see section 15.2.2). MP2 grade was probably somewhat lower than MS2.

Temperatures may have fallen further by D3 - MS3 garnet was not recorded, strained quartz has generally not been able to recover, and F3 crenulations have not been as thoroughly recrystallised as F2. However, biotite was stable, and plagioclase, while sometimes recrystallised at its margins, remains about An_{20}.

MP3 muscovite and biotite porphyroblasts need not indicate an increase in temperature, only a steady state or slow cooling as deformation ceased.

D4 deformation brings in the first retrogression, with chlorite replacing muscovite and biotite in typical pelites, coupled with marked reduction in quartz grain size.

D5 and D6 deformation also causes retrogression of garnet and biotite to chlorite, and quartz (±chlorite±calcite) veins become abundant. Chlorite and rare white mica porphyroblasts occur, while micas in F5 crenulations are highly strained and broken.

Timing of mineral growth in typical pelites and inferred qualitative temperature variation are shown schematically in Fig.3.6.

3.4.5 Discussion

The stable mineral assemblages can conveniently be split into three groups for discussion:-

i) the MP1-MS2 metamorphic peak (garnet grade)
ii) MP2-MS3 periods (biotite grade)
iii) MS4-MS6 periods (chlorite grade)

i) MP1-MS2

The stable assemblages here are quartz-biotite-muscovite-plagioclase(An_{20})-garnet-ilmenite(-calcite) in pelites, and quartz-plagioclase(An_{30})-biotite-garnet-zoisite-calcite in calc-silicates.

The pelite assemblages place the rocks clearly in the garnet zone (garnets have only minor spessartine and grossular). Unfortunately, high-grade Moine rocks do not usually contain staurolite, so its absence cannot be considered diagnostic of low grade (cf. section 16.1). However, the presence of only An_{20} plagioclase, in rocks which contain calcite or clinozoisite, suggests temperatures below the staurolite zone (Winkler 1976); this is supported by temperature estimates of _c._ 520°C at 6.5kb (section 15.2.2(i)) - cf. Winkler's suggestion of 550°C for this boundary. The calc-silicates correspond to Kennedy's (1949) calcite-zoisite zone, which he correlates with the Barrovian garnet zone; Winchester (1974) came to a similar conclusion.

ii) MP2-MS3

Assemblages formed in pelites during this period were similar to those in MS2, but with the absence of garnet, i.e. biotite, muscovite, An_{20} plagioclase were all stable, with abundant microcline in suitable rocks. MP2 garnet could be stabilised at lower grade by high Ca or Mn - e.g. in psammites garnet is rare, and will probably scavenge most of the Mn in the rock, and in calc-silicates, it contains over 30% grossular. The poikiloblastic garnets in calc-silicates, which may be MP2, give temperatures of c. 460°C at 5.8kb (section 15.2.2(i)); MP2 was probably at or just below the garnet zone boundary. The absence of garnet and poorer recrystallisation in MS3 may indicate somewhat lower temperatures, although An_{20} plagioclase was still stable.

If, as suggested by published radiometric dating (section 14.4), D1 and D2 are Precambrian and D3 and D4 Caledonian, this metamorphic history seems to suggest steady (albeit very slow) cooling over 500Ma (Grenville) or 250Ma (Morarian) until the onset of Caledonian reworking, which itself involved little reheating. This contrasts with Tobisch et al.'s (1970) suggestion of a metamorphic low separating early and late tectono-metamorphic events in the Moines. Such a pattern may only be recognisable in areas where the Caledonian event was of significantly higher grade than the latest widely-developed stages of the earlier metamorphism. In this area, close to the Moine Thrust, rapid post-Grenville uplift and early-Caledonian burial might not allow detectable post-MP2, pre-MS3 chlorite-grade assemblages to form, even if they could survive a later biotite-grade metamorphism. The critical evidence would be a chlorite-grade "D4" zone which was cut by a biotite-grade D3 zone. This was never observed, but since only six S3/S4 intersections were recorded, the possibility cannot be ruled out.

iii) MS4-MS6

All these events can be generally described as chlorite grade - chlorite replaces biotite, muscovite and garnet, and is common in

D5 and D6 veins. D4 quartz deformation was essentially plastic and D6 essentially by pressure solution; D5 shows some features of both. This may indicate declining temperatures, although decreasing strain rate could have a similar effect.

3.5 General Discussion

3.5.1 Comparison with literature

The conclusions regarding the peak of metamorphism, and the mesoscopic identification of S1 and S2 fabrics, are in general agreement with published work: e.g. Powell (1966, 1974). The later deformations are more difficult to correlate. Powell (1974) gave a four-fold sequence, but Poole & Spring (1974) reported only three major phases of folds and cleavages or crenulations. None described structures directly comparable to D3 and D4 of this thesis - zones of high strain not directly associated with folding. Some of Powell's F3 folds have similarities in style and in the degree of recrystallisation of micas to the F3 folds described here; Powell's F4, with conjugate folds and crenulations, is similar to F6 of this thesis; the F6 Glasnacardoch Bay structure is shown by Powell (1974) as an F4.

3.5.2 Major structures

The major fold in the area is the tight, locally overturned, Morar Antiform. This has been variously described as F2 (Powell 1974), F3 (Poole & Spring 1974) or F4 (Winchester 1974). The apparent retrogression in the core (Lambert 1958), which Winchester explained by the folding up of an earlier retrogressive isograd, is difficult to reconcile with an F2 origin, synchronous with the metamorphic peak. In Ardnamurchan, O'Brien (1981) believes the (en echelon) continuation of the Morar Antiform to be late, and the similar Sron na Gaoithe structure is at least an F3 (section 6.2.1). Vergence of parasitic F2 folds at Mallaig is consistent with an F2 Morar Antiform (Fig. 3.1); however F2 at Inverie, on the eastern limb, has the wrong vergence (section 4.5), and a traverse across the core of the antiform past Bracora encountered folds identical to F2 but with gently north-dipping axial planes, which were folded by open, upright structures. At Druimindarroch in South Morar, both bedding and S2 are gently-dipping in the core of the upright Morar Antiform (cf. Smith & Harris 1972). Where F3 vergence can be obtained, it is opposed to the Morar Antiform at Mallaig, but consistent at Inverie; platy zones of F3 aspect are gently north-dipping in the core of the major antiform. The Morar Antiform is therefore at least post-F3 (of this thesis). Three possibilities can be considered:-

i) F4 - no minor folds of Morar Antiform age are present, but the S4 shear zones are of this age. This model would be analogous to that of O'Brien (1981) in Ardnamurchan.

ii) F5 - the plunge of F5 folds is reasonably close to that of the Morar Antiform, but the axial planes would have to fan very strongly around the fold.

iii) no minor structures of this age are present at Mallaig - e.g. in the Assapol Synform on Mull (probably complementary), large segments of the limbs have been rotated with no development of minor folds or fabrics (section 7.2.4).

The third possibility is favoured by the fact that in Ardnamurchan, a grain size reduction zone similar to S4 cuts the Salen Slide fabric (probably equivalent to S3 of Mallaig), but is itself folded by a minor fold of the Sron na Gaoithe generation; however, the other two models cannot be positively ruled out.

Given that the Morar Antiform is F4 or later, S1 to S3 must once have been fairly flat-lying. By including data from Inverie (section 4.2), it can be seen that F3 fabrics are downward-facing on the western limb, and upward-facing on the eastern limb. F2 fabrics are anomalous, in that they face northwards on the western limb and southwards on the eastern limb. If the antiform is opened out, these become respectively westward-facing and perhaps reclined with E-W trending sheath folds - cf. Powell (1974), who shows early eastward-closing, Lewisian-cored isoclines in his fig. 5. F3 structures are probably synchronous with the Knoydart and Sgurr Beag slides - it is possible that F3 platy zones represent initially gently-dipping, westerly-directed thrusts. If so, the apparent rotations in Unit 3, which underlies a very intense platy zone (presumably representing a significant thrust) may be explained. It may be significant in this context that the Unit 4 pelites, overlying this thrust or slide, were traced south only 600m on the IGS Sheet 61.

CHAPTER 4

Inverie

4.1 Introduction

Inverie lies on the south shore of Knoydart, on the eastern limb of the Morar Antiform. The area was studied for two reasons: rocks usually correlated with the Morar Division are migmatised, which is unusual, and the migmatite front seemed to coincide with the Knoydart Slide, inviting comparison with the Sgurr Beag Slide elsewhere (Powell 1974, Poole & Spring 1974, Ramsay & Spring 1962).

Conventionally, the succession has been assumed to be essentially undisturbed, despite the presence of the slide. Thus the Barrisdale Psammite of Ramsay & Spring (op.cit.) was correlated with the upper part of the Lower Morar Psammite at Mallaig (cf. Chapter 3 and Fig. 3.1 - terminology of Johnstone 1975), and the overlying Ladhar Bheinn Pelite with the Morar (Striped and Pelitic) Schist of Mallaig. The presence of the slide and the metamorphic break associated with it (see below) casts some doubt on this stratigraphy; for the purpose of this chapter, a purely lithostratigraphic approach has been adopted. Three rock groups have been distinguished on Fig. 4.1: in ascending order, these are the Lower Psammite, Inverie Semipelite and Knoydart Pelite. Their characteristics are listed below.

iii) Knoydart Pelite: coarse grained migmatitic pelites with abundant 1cm lits and pegmatites up to 3m thick; becomes more semipelitic upwards. Rare calc-silicates present.
ii) Inverie Semipelite: very platy, non-migmatitic, muscovite-rich pelites and semipelites; base transitional to (i).
i) Lower Psammite: massive, cross-bedded grey psammites with abundant sedimentary structures; some semipelitic intercalations.

The Lower Psammite is continuous with that at Mallaig (I.G.S. Sheet 61 and Poole & Spring 1974), and is relatively undeformed, so this correlation is accepted. The non-migmatitic Inverie Semipelite probably corresponds to the lower part of the Morar (Striped and Pelitic) Schist - the Lower Striped Schist. Both show a transitional boundary with the Lower Psammite, and have the same metamorphic grade. The Knoydart Slide occurs at the boundary of units (ii) and (iii), and is marked by a break in metamorphic grade and very high strains; the relationship of the Knoydart Pelite to the other units is therefore debatable.

Two traverses were made (Fig. 4.1), one across the slide and one into the migmatites, mainly to assess the age of the slide, and the relationship of the migmatites to the slide.

4.2 Structure

The dominant fabric in the area is of F3 age (section 4.2.3) and is related to the Knoydart Slide. Earlier fabrics are seen in lenses and augen ranging from a few millimetres to a few centimetres in width. In the east, F3 structures open out and some earlier folds are seen. The two traverses will be described in terms of their orthogonal distance from the slide; the latter can be defined within 10m or so as the locus of the most intense platy fabrics, and the area in which non-migmatites give way to migmatites.

4.2.1 Northern Traverse

550 - 150m below slide

The rocks form part of the Lower Psammite and are essentially undeformed. Psammite beds are a few metres thick, with rounded pebbles and abundant sedimentary structures, including small-scale cross-bedding (Plate 4.1(a),(b)); younging is consistently upwards and eastwards. In the top 200m, thin pelite bands appear, carrying a relatively fine-grained, penetrative schistosity, which is similar in field appearance to S3 of Mallaig (section 3.3.2; (56137)). This fabric lies within about 5° of bedding (e.g. S-S1 25/108, S3 30/110) but is slightly steeper; intersection lineations pitch north or south at moderate angles (< 45°). Virtually all the deformation has been accommodated in the pelites.

150 - 50m below slide

Semipelites become more common and psammites are restricted to 10cm stripes. S3 is largely penetrative and dips at 40/108; psammites here are deformed and carry an LS shape fabric parallel to S3, with L component plunging at approximately 35/130.

50 - 0m below slide

The rocks are dominantly muscovite-rich pelitic schists with small garnets, and carry a strong S3 fabric dipping at about 35/110. S3 is lenticular in detail, so that traces of an earlier oblique fabric (probably S2) can be seen in some hand specimens (Fig. 4.2(a)). In thin section, lenticular relics of earlier fabrics and textures are seen, separated and truncated by finer grained zones of high D3 deformation (Plate 4.2(a),(b),(c)). Where the relations of this earlier fabric to bedding can be seen, it indicates an antiform to the west, with a steep southward plunge. As the slide is approached, the rocks become very fissile; S3 is totally penetrative for the last 30m, and intersection lineations pitch about 70°S, parallel to a strong mineral lineation.

Slide zone

The top of the Inverie Semipelite is marked by a zone about 20m thick in which the rocks are very fissile (1mm plates), and dip at 35/112; an intense schistosity carrying a mineral lineation pitches about 70°S, parallel to reclined folds of pegmatoid lits. Conspicuous 1-2mm muscovite porphyroblasts occur on the foliation, and ribbon-like quartz veins are common. Lithologies from both sides of the slide are finely interleaved, and all are highly deformed - e.g. migmatite lits are a few millimetres thick, with occasional porphyroclasts up to 1cm (Plate 4.1(c)).

0 - 50m above slide

The same textural features continue for 30m into the Knoydart Pelite, but in uniformly migmatitic rocks. Deformation there becomes heterogeneous, with zones of less deformed migmatite showing recognisable augen and fold cores, with lits a few centimetres thick (Plate 4.1(d)).

50 - 170m above slide

Migmatitic pelites are deformed but not platy - lits are 1-2cm thick, but carry a fabric; coarse lit quartz is recrystallised, while plagioclase is recrystallised and may be broken. Folds of lits and bedding are isoclinal and pitch 70°S on S3 (the latter dipping at about 40/120); intrafolial folds are seen in the S3 schistosity. Above 100m S3 is noticeably steeper than bedding.

170 - 400m above slide

D3 deformation decreases upwards - in many rocks, S3 is a strong crenulation cleavage, though F3 folds are still reclined. At 200m, the lithology becomes more semipelitic and schistose, with large (>2mm) garnets. These schists are poorly exposed, but it appears that the S3 fabric becomes lenticular and areas carrying largely S2 are preserved (e.g. 56168). S-S1/S2 vergence seems to indicate that a steeply south-plunging antiform lies to the west.

4.2.2 Southern Traverse

20 - 120m above slide

Highly deformed migmatitic pelites carry an S3 fabric parallel to banding (45/130), with a strong down-dip mineral lineation and quartz veining. The foliation becomes less penetrative in the upper part so that locally, garnets and lits are not augened.

120-190m above slide

S3 becomes oblique to bedding (e.g. S3 55/120, S-S1 30/102) and is a tight crenulation cleavage, axial planar to ENE-plunging folds (Plate 4.3(a)). Thin lits are transposed into S3, but thicker pegmatites preserve a cleavage/banding intersection; pegmatites up to 3 metres thick carry a strong S3 fabric.

190 - 300m above slide

The rocks are striped pelites and micaceous psammites, and F3 folds are relatively open, but the pitch of their axes varies from gentle to reclined (Plate 4.3(b),(c)). The S3 fabric is a strong crenulation cleavage of an earlier differentiated fabric (Plate 4.2(e)), and dips at about 45/110. S2 is oblique to bedding; the S2/S-S1 intersection still plunges south, but is much less steep, and indicates that an antiform lies to the east (Fig. 4.2(c)). In pelite interbeds, the development of the S3 layering by transposition of earlier migmatitic lits is particularly well seen (Fig. 4.2(d)). At the highest levels, F3 folds become open, although the crenulation can still be recognised. This overprints earlier interference patterns involving F1 and F2, in which the development of pegmatites and the migmatitic layering is seen to be MS2 - MP2 (Plate 4.3(d)).

4.2.3 Discussion

The two traverses have been combined to make the composite sketch section Fig. 4.2(e). The pattern of deformation is very similar to that found along the Sgurr Beag Slide (Rathbone & Harris 1979, Rathbone 1980), i.e. development of a strong platy fabric, elimination of planar discordances, and rotation of linear elements into parallelism with a strong mineral and quartz grain-shape lineation. These fabrics can be readily identified on either side of the slide, and provide a link between the migmatitic and non-migmatitic rocks. West of the slide, this is at least a second fabric, as shown by the lenticular relics in (56139). Textures (see below) such as the lack of annealing of strain features, style of the fabric itself and augening of garnet and microcline (the latter probably MP2 at Mallaig - section 3.4.2), make this fabric very similar to S3 at Mallaig. East of the slide, the fabric becomes less intense and folds appear, in the core of which a differentiated fabric similar to Mallaig S2 is crenulated; interference patterns also indicate an F3 age for the sliding. The pattern of deformation indicates a strain in which linear and planar elements increase in intensity together; a plane strain such as simple shear would fit with this (cf. Mull, section 7.4).

The orientation of S2 is difficult to define with any certainty within and below the slide zone. S2 in lenses a few centimetres thick is slightly steeper than S3, which appears to be parallel to bedding; lineations on S2 pitch very steeply SW, parallel to L3 in the slide zone. In less deformed rocks at Mallaig, S2 orientation was preserved in microlithons, so this could be used indirectly to define S2/S-S1 relations; however, here the strain is much higher, and the parallelism of L2 to the other lineations suggests that rotation towards the shear direction has taken place. If L2 was originally horizontal or south-plunging, then the original vergence would have been towards an antiform to the west. If L2 plunged northwards or eastwards at more than a few degrees (cf. the moderate northwards plunge at Mallaig, section 3.3.2), then the original F2 vergence would have been towards an antiform to the east, and the axes would have been rotated from NE pitches, through the reclined position to 70°SW. Where S2 can be clearly identified above the slide, it is shallower than bedding, and outcrop-scale F2 folds show an S-profile with a moderate southwards plunge.

4.3 Metamorphism

Two aspects will be considered: MS3 (syn-sliding) and earlier metamorphism.

MS3 metamorphism

The sliding is clearly lower-grade than the metamorphic peak - grain size is reduced, quartz, feldspar and micas are strained, garnet is augened and migmatitic lits are attenuated (Fig. 4.2(b), Plate 4.2(a)-(d)). New biotite and muscovite grow in the slide fabric, and plagioclase about An_{20} seems to be stable in the matrix of (56137) and (56165). As in Mallaig D3, micas in tight crenulations are recrystallised, but more open examples retain strain (Plate 4.2(e)). Muscovite porphyroblasts are common in the central few tens of metres, but these probably represent recrystallised matrix muscovite rather than metasomatic products. Metamorphism can thus only be described as biotite grade.

Pre-slide metamorphism

There is clearly a break in early metamorphic grade across the slide from schists to migmatitic gneisses. Although there is also a lithological change from the Inverie Semipelite to the lower part of the Knoydart Pelite (which is similar to the Garnetiferous Pelite of Morar), higher up in the migmatites, striped schists with muscovite pelites return; these are significantly coarser than those below the slide (micas and garnets several millimetres *vs.* 1mm, 1cm quartz lenticles).

i) West of the slide Here, pre-D3 matamorphic grade seems similar to that at Mallaig - pelites have quartz, garnet, plagioclase (An_{20}), biotite, muscovite, ilmenite, (± microcline, clinozoisite) - i.e. essentially indicating garnet grade.

ii) East of the slide Mineral assemblages here are basically similar to those to the west of the slide (due to the low Al, Mg content of the rocks, this assemblage is stable until muscovite + quartz breakdown - cf. section 16.1). However, plagioclase is more calcic (An_{25-30}) and clinozoisite is not present in pelites (cf. An_{20} plagioclase and clinozoisite in the similar Garnetiferous

Pelite at Mallaig). In addition, 1cm-thick lits with more feldspar than quartz are widespread, as well as pegmatites up to 3 metres thick. The pegmatites tend to be emplaced parallel to F2 axial planes; they and the lits are clearly deformed by F3 but are never seen to be deformed by F2. Quartz veins with minor feldspar are deformed by F2 (cf. Mallaig, section 3.4.3), and garnets which are augened by S2 show similar textures and zoning to those at Mallaig (MP1-MS2). Other, clear garnets grow in the lits and are syn- to post-F2. Inclusion-free garnets overgrow healed F2 crenulations in muscovite schists of the Knoydart Pelite (Plate 4.2(f)). It may be, therefore, that the Knoydart Pelite has suffered a similar MP1 metamorphism to the Morar rocks, but during and after D2 it was raised to higher, migmatitic, temperatures. Metamorphic grade can only be said with certainty to lie between garnet and sillimanite-orthoclase. The relatively calcic plagioclase and lack of epidote (cf. section 3.4.4) suggests grade comparable to staurolite or kyanite zones, as does the presence of migmatitic lits; the absence of K-feldspar in lits whose host contains muscovite suggests a sub-solidus origin (section 17.1). Tanner (1976) has compared calc-silicate assemblages elsewhere in the Knoydart Pelite with the kyanite or staurolite zones.

4.4 Conclusions

The Knoydart Slide is a relatively late zone of high strain, which post-dates the main metamorphism in both migmatites and non-migmatites, and juxtaposes the two across a narrow zone. Temperature estimates from west of the slide are similar to those from Mallaig (513°C, 5.7kb); the best estimate east of the slide is probably around 580°C at 6-7kb (sections 15.2.2(ii), 15.2.3(iii)).

The main published synthesis of the area is that of Poole & Spring (1974). They date the Knoydart Slide as F2 rather than F3; they also place it at the base, rather than the top of the Inverie Semipelite. F2 and F3 vergence essentially agree in the Knoydart Pelite (Fig. 4.2(f)); the reclined nature of L2 and L3 at and below the slide (cf. variable above it) is clearly due to the high D3 strain. Powell (1974) agrees with Poole & Spring on the age of the slide, but disagrees with their major folds.

Following the reasoning in section 3.5, that the Morar Antiform is F4, three models for the Knoydart Slide could be considered:

i) The slide is simply due to high strain on the common limb of the Morar Antiform and Poole & Spring's Ben Sgriol Synform; the slide is F4 and no Mallaig F3 structures are developed here.
ii) The slide compares with the F3 structures at Mallaig and is folded by the Morar Antiform.
iii) The slide is an F4 or later structure but represents thrusting of the Knoydart Pelite over the eastern flank of the earlier Morar Antiform.

These alternatives are sketched in Fig. 4.2(g). The scale of movement on the slide is clearly important in distinguishing these possibilities. If the thermometry is correct, then pre-sliding, there was a temperature difference of <u>c.</u> 65°C between the rocks on either side. This probably represents at least 3km vertically in a relatively high pressure terrain like the Moines, on a steep shear zone. If the barometry is reliable, the Knoydart Pelite was originally no more than 1-2km deeper than the rocks to the west, while the slide zone itself is of the same order of magnitude as the Sgurr Beag Slide, which probably has tens of kilometres displacement (section 7.4). These figures clearly favour models (ii) or (iii). If (ii) is correct, the slide might surround the closed outcrop of the Knoydart Pelite, and could link up with the Sgurr Beag Slide, or could be a lower branch. It would, however, have to come down in the west beyond Mallaig. Model (iii) avoids this requirement, but if the barometry is correct, is difficult to reconcile with the low pre-sliding pressure differential.

In either case, accepting the broad correlation of Morar and Knoydart metamorphism seems to imply a Precambrian age for both. The 751Ma Rb-Sr muscovite date from the Knoydart mica mine (Long &

Lambert 1963) could not have survived high-grade MP2 metamorphism, though it could have survived biotite-grade metamorphism during a Caledonian D3.

CHAPTER 5

Kinlochourn

5.1 Introduction

Kinlochourn is situated at the head of Loch Hourn (which separates Glenelg from Knoydart), and lies on the boundary between the Morar and Glenfinnan divisions. It was here that the Sgurr Beag Slide was first defined by Tanner (1971) as a major tectonic break separating the two divisions. Migmatisation is widespread in the Glenfinnan Division rocks, but in the Morar Division is restricted to relatively late, but syn-tectonic, pegmatite veins. A structural succession comprising four major lithological units can be recognised (cf. Fig.5.1):

iv) Migmatitic Pelite
iii) Migmatitic Psammite
(Sgurr Beag Slide)
ii) Semipelite
i) Psammite

i) Psammite This consists of massive, unmigmatised, quartzose psammite, often micaceous, containing bands and lenses of quartzose semipelite. It is locally cross-bedded and contains calc-silicate lenses. It is dominated by N-S trending upright structures, in the cores of which the rocks are right-way-up, dipping south. These are probably related to the Sgurr nan Eugallt fold (Fig.5.1); since the unit (ii) semipelites surround this core just below the Knoydart Pelite, it is likely that the semipelites lie stratigraphically above these psammites.

ii) Semipelite West of the slide, these are homogeneous, non-migmatitic, muscovite-rich semipelitic schists with some pelites. Small red garnets are present, and quartz is dominant over feldspars. Similar lithologies occupy an antiformal fold core east of the slide. Quartz veining is common, but migmatisation is essentially late and structurally localised.

iii) Migmatitic Psammite This is separated from the semipelites by a zone of high deformation - the Sgurr Beag Slide - so cannot be assumed stratigraphically to overlie them. It consists of highly migmatitic, feldspathic psammites with some quartzites and rare semipelites. Migmatisation is early in that it is deformed by syn- and post-slide structures, and is generally of lit-par-lit form. Some unusual lithologies are also present - feldspathic augen gneiss with microcline and plagioclase porphyroclasts, and spatially associated hornblendite. These occur near the junction of the psammites with the overlying pelites, and the hornblendites at least may represent Lewisianoid inliers.

iv) Migmatitic Pelite This consists of coarse, lit-par-lit pelitic gneiss, with subequal muscovite and biotite, large garnets, and quartz-oligoclase lits ranging in thickness from a few millimetres to a few centimetres. Some semipelite or micaceous psammite stripes are present.

Units (i) and (ii) are continuous with Ramsay & Spring's (1962) Barrisdale Psammite (equivalent to the Lower Psammite of Morar), but Tanner (1971) believes them to be a local facies of the lower part of the Ladhar Bheinn Pelite - in either case, they lie in the lower half of the Morar Division succession. Units (iii) and (iv) correspond, respectively, to Tanner's Reidh Psammite and Sgurr Beag Pelite, the lowest members of the Glenfinnan Division succession, which rests on the Sgurr Beag Slide.

5.2 Structure

Two major features dominate the structure of the area: the zone of high deformation corresponding to the Sgurr Beag Slide, and a series of steeply-plunging upright folds whose axial planes strike slightly east of north. These will be designated $D2_k$ and $D3_k$ respectively, and will be described first, followed by a discussion of earlier structures.

5.2.1 Slide deformation ($D2_k$)

Structural and stratigraphical evidence agree that the trace of the Sgurr Beag Slide runs along the boundary between the semipelites and migmatitic psammites in Fig. 5.1 (Tanner 1971). A distinctive, intense planar fabric is present in the vicinity of the slide. It comprises a very strong schistosity in pelites, a strong platy fabric with quartz ribbons in psammites and a strong schistosity in which lits are augened and quartz ribboned in the migmatites (cf. Rathbone & Harris 1979, Rathbone 1980). This fabric is presumably related to the sliding, so structural dating of the fabric dates the slide movement. The fabric decreases in intensity both to the west and to the east of the slide. To the west, the psammites are platy up to about 150m from the slide, after which the fabric cannot be distinguished from that normally developed on $F3_k$ limbs. Cross-bedding is not seen until more than 300m west of the slide. To the east the platy fabric is folded within 50m of the slide by $F3_k$ structures, but can still be recognised. It dies out rapidly but heterogeneously into the pelitic gneiss, along the $F3_k$ axial traces; quartz ribbons (Plate 5.1(a), (b)) are present in only the lower 100m of the pelite. The fabric decreases in intensity inwards, on all three sides of the pelite outcorp west of Kinlochourn farm (Fig. 5.1), and is intense throughout the outcrop of migmatitic psammite. A similar fabric goes round the $F3_k$ antiformal dome of semipelite south-east of the farm, although here it is overprinted by a strong $D3_k$ crenulation cleavage. This pattern suggests that the closed outcrop of semipelite is an inlier of the lithology immediately underlying the slide.

In detail, the $D2_k$ schistosity is usually penetrative, but by 150m west of the slide an earlier schistosity is present, and an intersection lineation is developed. In the migmatitic pelite, $S2_k$ is locally a crenulation cleavage (Plate 5.2 (a), Fig. 5.4(a)) and post-dates both regional migmatisation and the bulk of garnet growth. Reclined isoclinal folds appear at 50m from the slide, becoming more open with increasing remoteness, $S2_k$ being axial planar. These are probably $F2_k$ but some may represent earlier folds which have been tightened up (Plate 5.1 (c), (d)); some, in a calcareous psammite, appear to have syntectonic garnets (Fig. 5.4(b)).

Throughout much of the pelitic gneiss east of Kinlochourn farm, $S2_k$ is weak or absent, so that the first post-migmatitic structures are of $F3_k$ age, but locally a strong $S2_k$ fabric is present (comparable to that found 50-100m above the slide), over an outcrop width of a few tens of metres.

5.2.2 Post-slide structures ($D3_k$)

All the N-S post-slide folds have been called $F3_k$; although they vary markedly in style across the area, they are not seen to overprint one another.

They can be considered in four areas: (i) west of the slide; (ii) the complex synform east of the slide; (iii) the antiform south-east of the farm; (iv) structures further east.

i) West of the slide The structure consists of N-S deformed zones, and an E-W (generally gently SSW-dipping), gently folded, weakly deformed zone in which cross-bedding is preserved. The turnover is very sharp and it was not possible to determine the overall vergence of the structure. Further west, about easting 94, deformation in N-S trending rocks becomes heterogeneous, and alternate zones are platy, with reclined isoclinal folds of pegmatites, and weakly deformed, with discordant pegmatites (Fig. 5.4(c)). These pegmatites are part of the suite which ranges from pre- to post-$D3_k$, and are coarse, locally with graphic textures; their quartz-albite-microcline compositions contrast with the quartz-oligoclase regional lits. This suggests that much of the platy fabric in this area is of $D3_k$ age, developed on the limbs of the major structure. It is likely that some $D3_k$ strain is present on the eastern limb, i.e. in the slide zone at the lochside, so that the precise strain pattern here is not typical of the Sgurr Beag Slide regionally. This may explain the narrowness of the highly deformed zone here when compared with Glen Shiel (Rathbone & Harris 1979) and Mull (section 7.4). Fig. 5.4(d) shows how the width of a shear zone, defined by a certain strain threshold, may be reduced by a suitable superimposed homogeneous strain.

<u>ii) The synform core</u> The core is fairly open (interlimb angle about 60°), plunges about 60°S (Fig. 5.1), and contains a series of variable - amplitude M and W folds with an axial planar crenulation cleavage (Plates 5.1(c), 5.2(b), (c)). As the limbs are approached, the folds become tight and asymmetric, with highly deformed long limbs (Plate 5.1(e)). By [948066], the folds have the style shown in Fig. 5.3(b), similar to the major S-fold, and the crenulation cleavage has fanned so that here it trends NE-SW. On the hilltop at [947065], the $S2_k$ fabric weakens and becomes heterogeneous, and the pattern of folding and nature of axial planar and folded fabrics is identical to that seen in area (iv), suggesting that the dominant folds in that area are also $F3_k$ (rather than $F2_k$).

Moving east across the migmatitic psammite, $F3_k$ folds tighten and the axial planar fabric becomes penetrative, returning to an 015 trend. An LS shape fabric appears in the psammites (axial ratios approximately 9:3:1), parallel to $S3_k$ and with L pitching steeply south.

<u>iii) The antiformal dome</u> The closed outcrop pattern of the semipelite is well defined on the eastern, southern and south-western margins. In the north and north-west, the precise boundary is concealed by drift, but the general pattern is confirmed by minor folds within the semipelite; the boundary has been drawn to correspond with the I.G.S. 6-inch sheet 78SE (Inverness). The minor folds are tighter than $F3_k$ folds to the west, with a semi-penetrative axial planar fabric, but the steady eastward tightening across the migmatitic psammite makes it likely that they are of the same generation. In the southern part of the semipelite, $F3_k$ folds are congruous with those to the west, i.e. south-plunging S-folds pass east through M and W profiles to south-plunging Z-folds. In the central region, $F3_k$ plunge is variable to north or south, but vergence also changes systematically - e.g. south-plunging S-folds and north-plunging Z-folds in the west - indicating that fold axes are curving through the horizontal. In the north, $F3_k$ plunges become steep to the north, and minor fold patterns indicate the presence of a northward

closure as shown on Fig. 5.1. In the extreme north-east (in the map-scale Z-fold at [953066]), fold plunges pass through the reclined position to pitch steeply south, although profiles remain consistent. All these minor folds have characteristic $F3_k$ style, fold a strong platy fabric and earlier isoclinal folds, and cause rodding of previously ribboned quartz veins. The major structure is a very tight antiformal dome or sheath fold (Quinquis et al. 1978), with western limb vertical, eastern limb dipping 85° east; it plunges about 80° south at the northern and southern ends, passing through the horizontal in the central portion. This is sketched in Fig. 5.3(a), and is very similar to the Sguman Coinntich structure, which folds the probably northward continuation of the Sgurr Beag Slide in Ross-shire (section 8.2).

iv) The eastern area Immediately east of this antiform, the $F3_k$ structures in the migmatites are continuous with those in the semipelite. Beyond [954064], $F3_k$ structures become very complex, with both S- and Z-profiles being common. Folds are largely, but not always, steeply plunging and numerous Type 1 and Type 2 interference patterns occur. The outcrop pattern of the I.G.S. 6-inch map also shows complex interference patterns. $F3_k$ folds are tight, with a penetrative axial planar schistosity in pelites and an LS shape fabric in psammites. Fold style is variable (Plate 5.1(f), Fig. 5.3(b)), but always with highly attenuated limbs. Three factors confirm their $F3_k$ age: the continuum in fold styles from undoubted $F3_k$ folds east of the slide, the similarity of folds of early quartz ribbon fabrics to $F3_k$ folds in area (ii), and the fact that migmatite banding is affected by interference patterns - i.e. these are the second post-migmatisation folds.

5.2.3 Pre-slide structures

The only evidence here for pre-slide structures is the presence in the Morar Division of an earlier foliation oblique to bedding, of a crenulated fabric in the migmatitic pelite (Plate 5.2(a)), and of the deformed migmatitic banding throughout the Glenfinnan Division. In addition, some of the folds in the slide zone may be tightened-up earlier folds. However, the fact that the main metamorphism in both divisions pre-dated the slide makes it likely that earlier structures were more abundant, but cannot now be recognised due to the high $D2_k$ and $D3_k$ deformation.

5.3 Metamorphism

Again, it is convenient to deal separately with pre-sliding, and syn- to post-sliding events.

5.3.1 $D2_k$ and $D3_k$ metamorphism

Both these events produced a penetrative muscovite-biotite schistosity, and micas in crenulations are completely annealed, so they were probably at least biotite grade (Plate 5.2(a) - (e)). No $MS3_k$ garnets were seen, but some possible $MS2_k$ garnet occurs (Fig. 5.4(b)). Pegmatite emplacement, including locally derived veins and pods, reached a peak during and after $D3_k$, which might indicate an $MS3_k$-$MP3_k$ maximum. The degree of recrystallisation and the fact that veins are rich in An_{25-30} plagioclase suggest similar grade to the rocks east of the "Caledonian migmatite front" in Ardnamurchan (section 6.4), so lower to middle amphibolite facies (garnet to staurolite grade) might be a reasonable estimate. A widespread feature in all rocks is reverse zoning of plagioclase from about An_{27} to An_{33}. Where plagioclase is recrystallised due to $D3_k$ deformation, the reverse zoning occurs in each new grain, i.e. it must be an $MP3_k$ feature (Plate 5.2(f)). If the plagioclase was in equilibrium with garnet or epidote, this would indicate increasing T or decreasing P - i.e. it is consistent with an $MP3_k$ metamorphic high (cf. section 15.2.3(iv)).

5.3.2 Pre-slide metamorphism

West of the slide Small garnets in the semipelites west of the slide are always augened by $S2_k$ (Plate 5.2(e)); pre-slide veins are essentially quartz with minor feldspar; matrix plagioclase is about An_{25}. Migmatites are not developed, and the $S1_k$ fabric is a penetrative differentiated schistosity very similar to S2 at Mallaig (section 3.3). This all suggests similar grade to the

Morar Division regionally, i.e. garnet grade or lower amphibolite facies. A P,T estimate of 590°C at 5.5kb is somewhat higher than at Mallaig (section 15.2.2). The above comments also apply to the semipelite inlier east of the slide.

East of the slide Garnet is, in general, augened by $S2_k$ and $S3_k$ (Plate 5.2(a)-(d), and appears in some cases to be $MP1_k$. Pelites thus show assemblages of quartz-plagioclase(An_{30})-garnet-muscovite-biotite-ilmenite. This could indicate anything from garnet to sillimanite grade, since Moine compositions do not usually develop staurolite or kyanite (cf. section 16.1). The coarse grain-size of micas and garnets relative to Morar Division rocks (several millimetres vs. ½mm), the relatively calcic plagioclase, and the development of true migmatitic lits suggest at least middle amphibolite facies conditions, e.g. staurolite-kyanite zones. Tanner (1971) has reported kyanite-fibrolite pods from this area, and P,T estimates of c. 640°C at 6kb are in agreement with this (section 15.2.3(iv)).

5.4 Migmatisation

Two categories of migmatisation can be considered: pre- and post-slide. The former is represented by the regional migmatites of the Glenfinnan Division, the latter by local remobilisation, usually structurally controlled, and coarse pegmatite dykes.

5.4.1 Early migmatisation

This is the typical lit-par-lit gneiss of the Glenfinnan Division, the "permeation gneiss" of the Geological Survey. Lits consist of An_{25-35} plagioclase (of the same composition as the host rock) and quartz, with minor micas and garnet, and are clearly intensely deformed in the slide zone (Plate 5.1(a),(b)). The lits were originally very coarse (Plate 5.2(g)), with tabular feldspars, and probably developed by in situ segregation of leucocratic material. The absence of potash feldspar, correspondence of host and lit plagioclase composition, and relatively low temperature estimates all point to a sub-solidus origin (see also section 17.1).

In the cores of $F3_k$ folds at [95500641], where $F2_k$ deformation is minimal, the lits are very coarse (several centimetres), show irregular boundaries with septa of palaeosome, and seem essentially undeformed (Fig. 5.3(d)). This suggests that post-migmatisation, pre-slide deformation was negligible, and so migmatisation was probably $MP1_k$. This may partly explain the lack of $F1_k$ structures - i.e. they have been obliterated by extreme $MP1_k$ coarsening of the fabric. This migmatisation is sharply bounded by the slide, and contrasts with the quartz veining of similar structural age in the Morar Division; thus here, as at Inverie (section 4.4), a slide has juxtaposed rocks of differing initial metamorphic grade.

5.4.2 Later migmatisation

This is present throughout the area, in both Morar and Glenfinnan divisions. Two types can be distinguished. The first consists of quartz-plagioclase veins, developed in $F3_k$ axial planes, and usually carrying an $S3_k$ foliation. Their composition reflects that of their host rocks, and they have probably developed by segregation *in situ*; they are comparable to the quartz veins developed at lower grade elsewhere in the Morar Division (e.g. section 3.4.3). Some of the veins folded by $F3_k$ west of the slide may also be of this type. In the Morar Division psammites (where undeformed), irregular veins, pods and segregations of pegmatitic aspect occur; these too are probably of local derivation. The second group consists of coarse, massive, quartz-albite(or oligoclase)-microcline veins, some of which show graphic texture and are very similar to the members of the Lochan Coire Shubh pegmatite complex to the east (Chapter 10). These have probably been emplaced as dykes, and span a range of ages from pre- to post-$D3_k$ (Fig. 5.4(c), Plate 5.1(g),(h)). Some are folded on axial planes parallel to $S3_k$, but fold amplitude and position of major axial planes does not correspond to that of folds of bedding (Fig. 5.3(c)). These folds could be late $D3_k$, or due to a later E-W shortening which has only caused a homogeneous strain in the host rocks. All these phenomena extend at least as far west as Barrisdale. The first type is probably a genuine indicator of metamorphic grade (cf. section 6.4). The dyke-like pegmatites are more indirectly related - in all areas studied, they do not extend beyond the limits of the first type, but are often absent entirely.

5.5 Discussion

The geological history can be summarised as follows: the Morar Division was metamorphosed to lower amphibolite facies, the Glenfinnan Division to middle amphibolite facies, with migmatisation, possibly corresponding to the MS2-MP2 event at Mallaig and Inverie. The two were then juxtaposed by movement on the Sgurr Beag Slide; at least locally, no other major structures of this age were developed. The present outcrop pattern was then produced by a series of tight N-S structures, with essentially coeval remobilisation and pegmatite emplacement over the whole area. The slide itself could be related to a major fold, a steep shear zone or a thrust; the rotation of linear elements into a steep attitude suggests essentially up-dip movement (in its present orientation). The present steep dip is due to $F3_k$ reorientation, so the slide could initially have been gently dipping (Fig. 5.2). $D3_k$ structures show some variation across the area (Figs. 5.1, 5.2). In the west, steeply-dipping, highly deformed, and gently-dipping, undeformed areas alternate, with no obvious relationship to major folds. Immediately east of the slide, this gives way to complex folding, which becomes tighter with curvilinear axes further east. This may indicate that some deep-seated structure related to the slide is controlling the style of later deformation, or it may simply be due to the contrast in lithology between the dominantly psammitic Morar and pelitic Glenfinnan divisions. Although, in the east, more of the section is deformed, and steep plunges are more widespread, it need not follow that bulk $D3_k$ strain is higher. Essentially, the cores of the major structures become more deformed towards the east - but since very platy zones become less common, the limbs may be proportionately less deformed. The overall pattern in the vicinity of the domal structure could be explained by an interference pattern (Type 1($\rightarrow$ 2) of Ramsay 1967) - see Fig. 5.3(e). However, small scale patterns of this type are rare (though they become common further east), and since the slide ($D2_k$) is folded, it would require an additional post-$D2_k$, pre-$D3_k$ phase of deformation for which there is no independent evidence. The structure could readily be produced if the curvilinear nature of minor $F3_k$ axes is duplicated on a large scale. This could be caused by differential movement of material along the axial planes (Ramsay 1962) or a superimposed homogeneous strain (with $\underline{x}>\underline{y}>\underline{z}$, $\underline{x}$ steep) exaggerating an earlier slight curve in the $F3_k$ axis (cf. Sanderson 1973). Some elements of both mechanisms may be present.

The only published work on the area is that of Tanner (1971). There are some differences of interpretation, particularly as regards the age of the slide. He argued that the slide formed early in D2; however, major structures of his F2 age are here attributed to $F3_k$ (folds in the core of the Sgurr nan Eugallt structure) as are his F2 minor structures east of the slide. The slide fabric is clearly folded by $F3_k$ structures and an $S3_k$ crenulation cleavage develops from the slide fabric, so it is considered here that there is sufficient evidence to define a separate episode. Tanner's F1 probably corresponds in part to $F1_k$ of this thesis, in part to $F2_k$ - he mentions the presence of two pre-F2 phases west of here, on the north shore of Loch Hourn. The complex synform at [946068] is shown by Tanner as an F1 structure, but the clear axial planar crenulation fabric makes this untenable; the nature of the semipelite inlier southeast of the farm is also different, the outcrop pattern in this thesis agreeing more closely with that in the I.G.S. 6-inch sheet 78SE (Inverness). Tanner (op.cit.) considered migmatisation to be essentially D2 in age ($=D3_k$) and to extend across the slide - this corresponds to the later migmatisation of this chapter. In Tanner (1976), he describes an earlier migmatisation restricted to the Glenfinnan Division, which is reworked by the later event; in this chapter, it is concluded that the dominant banding in the pelitic gneiss belongs to this ealier event, with little renewed <u>segregation</u>, although <u>tectonic</u> reworking is often strong. On a regional scale (section 14.3), $D1_k$ probably includes both S-S1 and S2 of the Mallaig sequence, so that the slide is of regional D3 age and the N-S structures D4.

CHAPTER 6

Acharacle

6.1 Introduction

Acharacle lies at the western end of Loch Shiel, on the boundary between Ardnamurchan and Sunart (Figs. 1.1, 6.1, 6.2). The area mapped comprises about 6 square kilometres and extends east from the A861 road between Acharacle and Salen and south from Loch Shiel. The ground is essentially low lying, reaching a maximum height of 173m on the prominent ridge of Sron na Gaoithe, which corresponds to the main outcrop of migmatitic pelite.

Little published work is available on the area - it is shown on the I.G.S. 1-inch Sheet 52, but forms the eastern edge of the area described by Butler (1965), and lies just off the maps published by Brown et al. (1970). The boundary between the non-migmatitic Morar Division and migmatitic Glenfinnan Division runs just west of the road (Fig. 6.2). It coincides with a zone of extremely high strain, which is probably continuous with the Sgurr Beag Slide - this will be called the Salen Slide.

Three main lithologies are present:- non-migmatitic psammite (Morar Division), locally-migmatitic psammite (Glenfinnan Division), and migmatitic pelitic gneiss (Glenfinnan Division). The last consists of at least two separate outcrop units - a thin pelite adjacent to the slide, and the main outcrop on Sron na Gaoithe. A structural succession is given below, reading outwards from the core of the Sron na Gaoithe Synform.

Glenfinnan Division

Lochan Breac Pelite This is a coarse, migmatitic pelite, rich in biotite, with abundant quartz-oligoclase lenses (1 x 10cm) and a matrix consisting of smaller lenses and micaceous foliae. Large pink garnets (up to 1cm) are common. It contains bands of coarse, non-migmatitic semipelite and rare pods and lenses (up to 10cm thick) of calc-silicate.

Lochan na Bracha Psammite This is a variable series of quartzites, feldspathic psammites and semipelites, with bands up to 20m thick of pelitic gneiss. Both psammites and semipelites often show lit-par-lit migmatisation; no sedimentary structures were seen. The upper few tens of metres are transitional to the overlying pelite.

Ruighe a' Phollain Pelite This pelitic unit consists of a structurally higher, non-migmatic, muscovite-rich pelite, and a lower migmatitic pelite similar to that of Sron na Gaoithe; both are highly deformed.

Morar Division

Carn a' Bhalbhain Psammite This consists of massive psammite with bands of non-migmatitic semipelite and rarer calc-silicate ribs. In this area it is highly deformed and no sedimentary structures are preserved.

This Morar Division psammite overlies a thick pelitic unit on the eastern flank of a probable en echelon continuation of the Morar Antiform (Figs. 2.1, 6.1). It is probably part of the Upper Moidart Psammite of Brown et al. (1970), equivalent to the Upper Psammitic Group of Morar (Richey & Kennedy 1939). This is separated from the Glenfinnan Division by the Salen Slide, so that their stratigraphic relations are uncertain. However, within the Glenfinnan Division, there are no high strain zones on the scale of the slide, so its internal stratigraphy is probably relatively undisturbed. In the literature, the pelites in Fig. 6.2 are generally correlated with the Lochailort Pelite (Brown et al. 1970, Johnstone et al. 1969), but this involves accepting all the psammites in this area as infolds of Upper Moidart Psammite. The high pre-slide metamorphic grade (section 6.3.1) and abundance of typical Glenfinnan Division lithologies makes this unlikely. Consideration of the structure, and reconnaissance further north, suggests that the main Lochan Breac Pelite may be equivalent to the Lochailort Pelite, while the underlying psammite and pelite represent lower structural levels in the Glenfinnan Division.

A short traverse eastwards from the slide in Salen Bay (Fig. 6.1) was also made, mainly to enable collection of fresh samples of slide rock. Lithologies are similar to those in the structurally lower of the Glenfinnan Division pelites, but thick garnetiferous amphibolites are present, probably representing early (pre-migmatisation) basic dykes.

6.2 Structure

The structure of the area is dominated by N-S folds, on both major and minor scales. The major folds are the north-closing Sron na Gaoithe fold and the south-closing Dig a' Bhogha fold (Fig. 6.2). The former plunges south at 60-80° throughout the area, but the latter has very variable plunge, and actually plunges northwards in the northern area. Minor structures of this age are widespread; these show the folds to be local F3, designated $F3_a$. The other major feature is the Salen Slide, which runs up the western boundary of the Glenfinnan Division and appears to be folded by the major $F3_a$ fold pair. It is most convenient to describe $F3_a$ structures first, followed by $F2_a$ (slide) structures, and then later structures.

6.2.1 $F3_a$ structures

Major structures The major $F3_a$ fold structure is the Z-profile fold pair formed by the Sron na Gaoithe and Dig a' Bhogha folds. These folds are tight with near-vertical axial planes (Figs. 6.2, 6.3, 6.4). Dip of banding on the western limb of the Sron na Gaoithe synform ranges from 70° east to vertical, with NNE strike; the eastern limb is generally steeper (always > 80°) with NNW strike. Since the axial plane must be nearly vertical, the dip of banding in the hinge is a good estimate of the plunge. This is 65-85° to the south at the northern closure [700672] and 50-60° to the south at the southern closure [698654]. Dip of banding in the core of the subsidiary antiform at [696651] is 80-85° to the south, and in a 100m-wavelength $F3_a$ fold NE of Lochan Ruighe a' Bhainne is 65-90° to the south. These data are summarised in Fig. 6.4. The maximum at approximately 85/110 corresponds to the western limb, the smaller grouping at 80/270 - 80/250 to the eastern limb. Its smaller peak reflects the fact that the eastern limb covers a much smaller outcrop area than the western (cf. Fig. 6.3). The grouping at 80/080 mainly derives from observations made in the platy zone on the eastern limb (see below), where $F3_a$ folds are isoclinal and axial planes dip steeply east. A weak girdle extends between the peaks representing the limbs, and indicates a plunge of approximately 70-75/190, on a vertical, 010-trending axial plane.

The style of the Sron na Gaoithe fold is markedly asymmetric. The western limb extends for 1km to the Salen Slide and has associated with it a large number of smaller folds; these are generally rather open in style, each with a relatively undeformed, south-dipping common limb (Fig. 6.3; map scale folds at [688666], [693662], [696670]; many folds of 20-50m wavelength in the Glenfinnan Division psammites). Some tight, symmetric folds also occur, e.g. the 100m wavelength synform to the east of the stream which flows from Lochan Ruighe a' Bhainne, and the major fold at [695652], which passes north into the open fold SE of Lochan Ruighe a' Bhainne. Minor structures indicate that these are also $F3_a$ folds. The eastern limb is narrower and more highly strained than the western, with $S3_a$ generally penetrative in the pelite, and with very tight minor folds in the psammite. In particular, the northeastern boundary of the pelite is affected by a 160-trending platy zone, 150m wide, with very intense $S3_a$. The spread about the vertical of the two peaks in Fig. 6.4 reflects that of measured $S3_a$ cleavage and is probably due largely to later folding.

The Dig a' Bhogha fold is a much more variable structure. Although fold profiles on the map are consistent with an overall Z-shaped structure (Fig. 6.2), the Dig a' Bhogha fold is a north-plunging synform in the north of this area. South of Loch Shiel, there is a broad core containing open, upright folds plunging north at 20-30°. Plunge steepens southwards and westwards within the same axial plane, while the folds become tighter and the rocks generally appear more deformed. The folds become reclined and isoclinal in a broad, NNW-trending platy zone, and the outcrop of the core (indicated by minor fold vergence) becomes narrower (Fig. 6.3). In the extreme southwest, the rocks are less deformed and folds plunge southwards. The two folds are thus linked by rotation of their axes through the vertical. The platy zone is somewhat oblique to $F3_a$ axial traces - e.g. Plate 6.1(a) was taken looking north from platy rocks at [710668]. The regional significance of these folds is discussed in section 6.5.

Minor structures The dominant minor folds are of this generation. They range from open to tight, depending on location (Plate 6.1(b)-(f)), and in the west often have the same open, asymmetric style as the medium-scale folds. The minor folds are essentially congruous on the major structures, but their plunge is more variable - e.g. on Sron na Gaoithe it ranges from 10° to 70° (Fig. 6.4). The folds generally have a strong axial planar crenulation cleavage in semipelites, a weak quartz shape fabric in psammites and a foliation developed by transposition of lits in pelitic migmatites (Plate 6.1(g),(h)). In fold cores, the crenulations are more open (Plate 6.2(a) - (c)), and fold an earlier crenulation cleavage (confirming that they are at least F3); no strain features are preserved in any of the micas (Plate 6.2(d)). In N-S striking $F3_a$ limbs, the foliation is penetrative, and augens individual plagioclase and garnet grains (Plate 6.2(e)). A characteristic feature of the $F3_a$ folds is the development of "platy zones", i.e. areas of apparently high deformation in which the $S3_a$ fabric is very strong and banding is strictly parallel, with few if any intrafolial folds (Plate 6.2(f)-(h)). These are usually developed on $F3_a$ fold limbs, but are not always present, and may be somewhat oblique to both regional bedding and $S3_a$ - e.g. the zone at [702670] cuts across folded bedding and must have formed late in $D3_a$. The rocks in the eastern platy zones show no distinctive microtextural features, probably due to $MP3_a$ annealing (section 6.3.2), but rocks in the west, which appear identical in hand specimen, contain numerous narrow zones of matrix grain-size reduction, which are very similar to the S4 zones at Mallaig (but contain biotite rather than chlorite - cf. section 3.3.2). These platy zones occur on the western limb of the Sron na Gaoithe Fold, and particularly in the Salen Slide zone where it trends N-S - there, they clearly cut the dominant $S2_a$ fabric.

The $S3_a$ fabric has a regular N-S orientation and is axial planar to the major folds: e.g. on Sron na Gaoithe it strikes from 080° to 110°, and dips range from vertical to 80°E (Fig. 6.4). Plunges follow the major folds, but are more variable; e.g. on Sron na Gaoithe, they range from 10-80°S, but are generally 60-70°S. In the eastern platy zone, folds are essentially reclined; to the north, fold axes become more shallow until they plunge north at about 20°, although variation from 10° to 40° may be seen in one exposure.

6.2.2 $F2_a$ structures

These include all structures which deform the migmatitic banding, but are themselves affected by $D3_a$ structures. The major structure of this age is the Salen Slide, a zone of high strain which marks the western limit of migmatitic Glenfinnan Division rocks. The strain pattern here is comparable to that at outcrops of the Sgurr Beag Slide (e.g. Rathbone & Harris 1979; section 5.2.1); reconnaissance further north and inspection of the 1-inch map suggests that the Sgurr Beag Slide continues as far as Salen (cf. Rathbone 1980). Strain, as indicated by attenuation of bedding, elimination of planar discordances and rotation of fold axes into a nearly reclined position, increases towards the boundary, although here cross-bedded psammites do not occur for several hundred metres to the west of the slide (probably due to superposition of $D3_a$ strain - cf. O'Brien (1981)). Psammites in the Morar Division are platy with ribbon-like quartz veins (Plate 6.3(a)), but in thin section show little indication of the strain that they have undergone. The rocks are completely recrystallised with unstrained, polygonal quartz grains, some strongly aligned, dispersed biotites, and cracked, strained feldspars with axial ratios of 2 or 3 to 1 (Plate 6.3(b)). Ribbon-like quartz aggregates occur in which all the grains extinguish within 10-20°

of each other; these are not suitable for detailed measurement but dimensions of 5mm x 0.25mm would be typical (in (58125), 30m west of the slide), and may give some indication of the strain. For example, if the strain was accomplished by an up-dip simple shear, and this was an estimate of the x:y ratio of the strain ellipsoid (since the thin section is virtually horizontal), shear strains of about 20 would be implied (cf. Mull, section 7.4). Such estimates are necessarily crude, but everywhere confirm that the strain is very large.

In the migmatitic pelite, lits and individual plagioclase and garnet crystals are augened by a penetrative foliation (Plates 6.3(c),(d); 6.5(a)). Considerable recrystallisation and coarsening of micas has followed this deformation, but quartz in the lits is reduced to less than 1mm and strained. The $S2_a$ fabric is penetrative throughout the pelite outcrop; at Salen, lits show very strong shape fabrics in quartz and recrystallised feldspars (e.g. 10:1 in feldspar in (56264), 30-40m east of the slide). About 100m east of the slide, reclined isoclinal folds of lits occur, and are probably of $F2_a$ age; some of these are refolded by $F3_a$ structures.

Elsewhere in the Glenfinnan Division, the dominant folded fabric is probably $S2_a$ or composite $S1_a$-$S2_a$. This varies in intensity from a crenulation cleavage with associated microfolds in the southern Sron na Gaoithe fold core (Plates 6.2(d), 6.3(e)) to a penetrative foliation which augens garnets and migmatitic lits in the centre of the pelite. A strong platy fabric which post-dates migmatisation in the psammites (Plate 6.3(f)) is folded round the main Sron na Gaoithe fold closure in the north. It forms a zone following the lithological boundary, about 50m wide in both pelite and psammite. Although narrower and less intense than the Salen Slide zone, it clearly represents a similar zone of high strain and possible displacement. Some strain features are preserved in these platy rocks - e.g. sub-grains occur in the quartz matrix, and quartz in the lits has clearly been recrystallised from several centimetres to less than 1mm (Plate 6.3(g)). Pre-$F3_a$ isoclinal folds appear within 20-30m of the lithological junction, and are present throughout the Glenfinnan Division psammites.

The main pelite outcrop on Sron na Gaoithe forms a Type 3 interference pattern (Ramsay 1967); the early fold core can be recognised by the presence of a zone of abundant fold closures, often quite open, which are overprinted by $F3_a$ structures (Fig. 6.5(a)). It is clear from the distribution of early folds, and the microstructures in the [695651] area, that the early fold core runs approximately WSW, south of northing 66, and that the folds between [695662] and [698664] are of $F3_a$ age (Fig. 6.3). Since the associated minor structures fold the migmatitic banding, this early fold is probably $F2_a$ in age; this is supported by the strain pattern described above. If the $F3_a$ fold is opened out, the platy (now northern) limb of the early fold would be nearest to the Salen Slide, while the present southern limb, which carries an $S2_a$ crenulation cleavage, would be furthest from the slide. The evidence is thus consistent with the early fold having been an isocline above, and possibly contemporaneous with, the slide.

The most common interference patterns are those involving $F2_a$ and $F3_a$, with migmatitic banding, as represented by coarse lits, going round both folds, although fine-scale banding may be transposed into one or both axial planes (Plates 6.1(a), 6.3(h)). Some more complex interference patterns occur - one shows a pre-$F2_a$ fold phase (Fig. 6.5(b)).

6.2.3 Earlier structures

No major pre-$F2_a$ structures were seen. Minor structures are restricted to rare interference patterns, e.g. Fig. 6.5(a),(b). An earlier foliation clearly existed, since $S2_a$ is a crenulation cleavage; the migmatitic banding, and the garnet grade metamorphism of the Morar Division predate $F2_a$. The style of the $S2_a$ fabric at the Salen Slide is very similar to that of the S3 fabric at Mallaig further north (section 3.3.2); reconnaissance further west shows that it overprints a strong, differentiated (i.e. striped or segregated) schistosity, which is often about 10° oblique to bedding, and whose field appearance is identical to that of the S2 fabric at Mallaig. Work by O'Brien (1981) has confirmed that this is a second fabric, i.e. that the Salen Slide is a D3 structure in a regional sense.

6.2.4 Late structures

Various late open folds and kink bands affect the $S3_a$ foliation - most are approximately east-west and vertical; they may in part be responsible for the variation in strike of $S3_a$ in Fig. 6.4. In the east of the area, within the zone of "Caledonian migmatisation" (section 6.4.2), two other types of late structure become common in psammites carrying strong $S3_a$. These are open folds which appear to form a conjugate set with NW-SE and NE-SW trends, and kink-like features (Plates 6.2(g), 6.4(a)) and shear zones indicating E-W compression (Plate 6.4(b)). These are similar to the late structures seen in more eastern parts of the Glenfinnan Division, e.g. Loch Quoich (section 11.2).

6.3 Metamorphism

The metamorphic history will be considered in two parts - pre-sliding, and syn- to post-sliding. Pre-slide metamorphic grade was essentially uniform within either Morar or Glenfinnan divisions, with a break at the slide, while syn- to post-slide grade varied uniformly over the area.

6.3.1 Pre-slide metamorphism

Morar Division There is little to define the precise grade in the small area studied. Garnet was stable in most rocks - i.e. psammites, pelites and calc-silicates; the last have andesine - biotite/hornblende assemblages generally taken to indicate something near the top of the Barrovian garnet zone (e.g. Kennedy 1949, Butler 1965). Fe-rich epidote is present in calc-silicates and psammites. Such features as coarseness, size and abundance of garnets, and degree of segregation are similar to the Morar Division elsewhere, suggesting garnet zone in a medium pressure facies series, although P,T estimates of 585-610°C at 6-6.5kb indicate somewhat higher grade than at Mallaig (section 15.2.2 (iii)).

Glenfinnan Division These rocks are coarse lit-par-lit gneisses, with lenses and segregations of quartz-plagioclase

(An_{20-30}) composition (Plate 6.4(c), (d)). Their coarse, migmatitic appearance suggests middle to upper amphibolite facies, i.e. staurolite zone or higher. The host rock consists mainly of coarse biotite, muscovite and garnet; the latter often contains coarse, irregular quartz, biotite and plagioclase inclusions, and by analogy with the rocks at Kinlochourn, is probably broadly coeval with migmatisation (section 5.3.2). Calc-silicates have anorthite (An_{92-95})-hornblende-sphene assemblages (e.g. 58116); these are retrogressed in the slide zone (56263). They indicate at least kyanite grade according to Kennedy (1949); although Winchester (1974) suggests that the calc-silicate isograds are sensitive to variations in the CaO/Al_2O_3 ratio of the rock, the main effect in this case is to allow anorthite-hornblende to correspond to kyanite <u>or</u> sillimanite zones. Pelitic index minerals are rare in these rocks; fibrolite is recorded at a number of localities on the I.G.S. 6-inch maps, but none was found in any samples collected for this thesis. Butler (1965) reported one occurrence of kyanite and staurolite. P,T estimates based on mineral chemistry suggest about 640°C at 6kb, which would lie close to the kyanite/sillimanite phase boundary (section 15.2.3(v)).

6.3.2 Syn- to post-slide metamorphism

<u>$MS2_a$</u> Metamorphic grade during sliding was clearly lower than before - garnets and migmatitic <u>lits</u> are augened, and high grade textures are broken down. However, biotite was still stable, and plagioclase which recrystallised during $D2_a$ remains high oligoclase. Metamorphism was at least biotite grade, probably quite high in the biotite zone, since plagioclase is stable relative to albite-clinozoisite. The degree of recrystallisation is greater than that of comparable structures at Mallaig or Inverie (sections 3.4, 4.3), but less than that at Kinlochourn or Sguman Coinntich (sections 5.3.1, 8.3.2), possibly indicating an intermediate grade; no new garnet appears to have grown.

<u>$MS3_a$-$MP3_a$</u> The main post-sliding recrystallisation is of this age. At least in the west, conditions were similar to $MS2_a$ - i.e. new biotite and muscovite grew, and strain features in quartz and

micas were annealed. There appears to be an eastward increase in grade - grain-size of rocks deformed in $D3_a$ increases from about 0.2mm to 0.5mm, and micas defining $S3_a$ become very coarse in the east (several millimetres). Zones of high (late) $D3_a$ strain in the west show grain-size reduction of the typical Mallaig S4 type, while in the eastern zones the quartz matrix is coarse and polygonal. Since $D3_a$ strain features are eliminated, this could reflect an eastward increase in $MP3_a$ grade. An "isograd" can be defined, running approximately N-S, just west of easting 71, and east of which the 2 or 3:1 $S3_a$ shape fabric is annealed to a coarse, equant texture (Plate 6.5(d); cf. Plates 6.2(c), 6.4(e)), plagioclase grains show slight reverse zoning in the outer few tens of microns, and "late migmatisation" (see below) becomes abundant. This line is shown as "Caledonian migmatisation" in Fig. 6.1; its significance is discussed in section 6.4.2. As at Kinlochourn, the reverse zoning may indicate $MP3_a$ increase in metamorphic grade (cf. section 5.3.1). It is suggested in section 6.4.2 that this represents something in the region of garnet or staurolite grade. The late structures in the east - conjugate folds and shear zones - show similar segregation phenomena and annealing, so the $MP3_a$ peak must have outlasted them. Various retrogressive effects were seen - e.g. patches of chlorite, sericitisation of feldspar, growth of white mica, calcite and prehnite in calc-silicates - particularly in the Salen Slide zone - but could not be related to any distinct tectonic event. It is likely that the slide, as a zone of fissile rocks at a major lithological boundary, has been reactivated under later stresses which have not caused regionally developed structures (e.g. that responsible for foliating some of the microdiorites in the area); in addition it may have acted as a pathway for fluids causing enhanced retrogression (essentially hydration and carbonation) in its vicinity.

6.4 Migmatisation

The term migmatisation is used rather broadly here to describe any high grade segregation phenomenon in which feldspar is present in significant quantities, i.e. >50%. Two types can be recognised on structural grounds - one which predates the sliding, and one which postdates it.

6.4.1 Early migmatisation

As at Kinlochourn (section 5.4.1), lit-par-lit migmatisation is widespread in the Glenfinnan Division; the lits are deformed by $F2_a$ and $F3_a$ structures (Plates 6.1(d), (h); 6.2(c), (f); 6.3(c), (f); 6.5(a)). The Salen Slide forms a precise western boundary to the pre-sliding migmatisation, a situation typical of the Sgurr Beag Slide throughout its length. The obvious conclusion, supported by the metamorphic evidence (section 6.3.1) is that the Glenfinnan Division was originally at a higher (migmatitic) grade than the Morar Division, and that they were juxtaposed by the slide after they had largely cooled. The early lits are generally trondhjemitic (i.e. quartz-oligoclase) in the pelites, which correspond to the typical Glenfinnan Division regional migmatites. Lits are typically a few centimetres thick and range from 10cm to several metres in length; they invariably lie parallel to the foliation, although this may be due to the high deformation throughout this area. The host rock is also segregated, on a scale of a few millimetres; this may represent thinner lits which have been transposed into the foliation. The relatively low grade (below sillimanite/K-feldspar) and absence of K-feldspar in lits whose host contains abundant muscovite make a partial melting origin unlikely (see section 15.3.1). Some lits do contain K-feldspar, but only in psammites where K-feldspar is already present in the host. Since the conclusion that early migmatisation is restricted to the Glenfinnan Division seems to have regional validity (section 14.1), the presence of early migmatites can be used to distinguish Morar from Glenfinnan psammites, and thus confirm that all the psammites south of Loch Shiel and east of the slide are Glenfinnan Division.

6.4.2 Later migmatisation

Broadly, this encompasses segregations which are syn- to post-$D3_a$. Little or no syn-sliding migmatisation occurs here, in contrast to areas further north such as Kinlochourn or Sguman Coinntich. This event is represented by veins, pods and lenses of quartzofeldspathic material. While it is appreciated that some of these, particularly the coarser pegmatite veins, may be distant members of vein complexes such as that at Lochan Coire Shubh (Chapter 10), and are essentially cross-cutting dykes of little local metamorphic significance, others have a more local derivation and clearly indicate something about the metamorphic grade of their host rocks. The first part of this section will concentrate on description, and later their significance will be discussed.

Veins which are broadly syn-$D3_a$ are seen in Plates 6.1(c), 6.4(f), (g). In the first case a vein is emplaced parallel to an $F3_a$ axial plane on a moderately deformed limb; it carries a very weak fabric, which contrasts with the strong shape fabric in earlier, folded lits (compare this case with the transposed early lits in Plate 6.1(h)). This vein is fine grained and ends when the limb becomes less deformed, and may be locally derived. The next example shows a ptygmatic vein and the third a vein which is folded but on different axial planes to bedding. Both of these are coarse, sharp-sided and K-feldspar bearing and probably are intrusive pegmatites. Plate 6.4(a), (h) shows segregations in psammites east of Sron na Gaoithe. The first is only slightly coarser than the host, has approximately the same composition and has diffuse boundaries with coarsened psammite which retains its original foliation. The second shows a similar coarsening, but has a composition similar to its host, minus its biotite; a biotite selvedge marks much of the irregular boundary. The shape of both of these (irregular, with opposing boundaries which do not match up) makes it unlikely that they opened as cracks, and an origin by segregation of the quartzofeldspathic portion of the host is more likely. The pod which has formed on the fold crest to the left of the hammer in Plate 6.4(a) may be of either local or distant origin. All of these are post-$D3_a$ (the foliation in the psammites is a platy $S3_a$). In Plate 6.4(b), coarsening and some depletion in mafics takes place in a late shear zone - this is the sort of feature which often develops into a pegmatitic segregation at Loch Quoich (section 11.3), and is presumably triggered by the deformation. Feldspar porphyroblasts in this example cut $S3_a$ and presumably are related to the same metamorphic event.

All these phenomena show a west to east zonation. In the west, they are scarce, and dominated by planar-sided pegmatitic veins and dykelets. In the east, the veins are more abundant, and diffuse patches and veins of coarsened material, often depleted in biotite, appear for the first time. Further west, similar structural positions are occupied by quartz veins with minor feldspar. Together with the evidence for increased annealing in the east (section 6.3.2), this suggests a west to east increase in metamorphic grade. An "isograd" can be defined where quartzofeldspathic (as opposed to quartzose) segregations become abundant; this more-or-less coincides with a noticeable coarsening, particularly of psammites. A contrast in the style of obviously intrusive pegmatites, from planar and sharp-sided, to irregular with crystals growing across the boundary, can also be seen - presumably indicating higher temperatures in the wall rocks during emplacement. The precise significance of this line in terms of temperature is difficult to quantify. Clearly from the textural evidence, $MS3_a$ in the west was at least biotite grade; since plagioclase was stable rather than albite-clinozoisite, a position high in the biotite zone, perhaps even into the garnet zone, seems most appropriate. Grade must have risen towards the east but by how much? One way of estimating the grade would be to compare $MP3_a$ textures here to those developed at known grades in other Moine rocks. East of the "isograd", feldspar is more abundant in segregations and matrix textures are much coarser than at Mallaig, where regional MP2 textures are preserved (garnet grade - about 520°C+, section 3.4.5). Rocks here which match the Mallaig rocks in these respects (dealing only with $MS3_a$-$MP3_a$ textures) would lie around easting 70 (Fig. 6.2). Other things being equal, this may suggest that those rocks were about garnet grade during $MP3_a$. Similar coarseness and vein composition to the eastern rocks are shown by the Knoydart Pelite (probably kyanite grade at approximately 580°C, section 4.3) - this seems a large temperature jump over 1km or so, however. In other metamorphic terrains, this coarse, segregated texture tends to appear about the staurolite zone; unfortunately, any staurolite zone Moines which might be used for comparison have been cut out by the Sgurr Beag Slide. Vidale (1974) recorded a similar pattern of vein composition in an apparently continuous metamorphic sequence, and found that plagioclase appeared in veins at garnet grade, but did not become common until staurolite grade. Thus the "isograd" probably corresponds to something in the upper garnet or lower staurolite zones. Since there is reason to believe (section 14.4) that the early migmatisation is Precambrian and the later one Caledonian, this line will be called the Caledonian migmatite front.

Two questions arise from the above interpretation of $MP3_a$ grade: why is there apparently no $MP3_a$ garnet, and why is the Caledonian migmatisation only locally effective? The first point could have a number of explanations. If the rocks were already saturated with garnet after $MP1_a$, and it remained refractory at these temperatures, no new garnet could grow. Alternatively, if the event was short-lived, garnet growth might be too slow to permit much crystallisation. Also, with or without both previous factors operating, garnet textural relations are not sufficiently clear to differentiate a narrow rim of $MP3_a$ age, due to the lack of clear inclusion trails and to $MP3_a$ coarsening of matrix micas (Fig. 6.5(c)). Mg-poor rims in some garnets may date from the $MP3_a$ event (section 15.2.3(v)).

Regarding the lack of new migmatisation, the pelitic gneiss, once fully segregated into quartzofeldspathic and micaceous portions, seems extremely stable - e.g. at Lochan Coire Shubh and Loch Quoich (sections 10.3, 11.3), later metamorphism reached partial melting temperatures, but pre-existing lits show no tendency to amalgamate or re-segregate, except when first broken down by deformation. Psammites, however, tend to be incompletely segregated in the early migmatisation, and usually retain bands of mixed quartz-feldspar-mica which will provide suitable starting materials for the production of new veins or lits.

6.5 Discussion

The conclusions so far can be summarised as follows: the Morar Division was metamorposed to garnet grade or higher; the Glenfinnan Division was metamorphosed to kyanite/sillimanite grade and migmatised. No structures of this age are preserved. They were then brought together by movement on the Salen Slide, with major isoclinal folding in the Glenfinnan Division. The present structure was then produced by upright, tight N-S folding, with variable plunge, and rising metamorphic grade towards the east.

This can be extended regionally by using the criterion of pre-slide migmatisation to define the Glenfinnan Division. All the ground south of Loch Shiel and east of the slide is occupied by undoubted Glenfinnan Division rocks (Fig. 6.1). To the north of the loch, a traverse eastwards from north of Acharacle crosses gently folded Morar Division psammites with sedimentary structures. At about Langal farm [710698], the first undoubted Glenfinnan Division lithologies are encountered - migmatitic pelite with lits transposed in $F3_a$-like structure (Plate 6.5(b), (c)). Further east, abundant lenses of migmatitic pelite and semipelite, along with migmatitic psammites and typical Glenfinnan Division quartzites confirm this identification. To the west, the exposure is poor, and the rock types are rather ambiguous. However, a zone of very platy psammites occurs at [706696], and probably marks the line of the slide. Further north, at Kinlochmoidart, the pelites at [708725] are typical early migmatites and are highly deformed (Plate 6.5(e)). Platy psammites occur just west of the farm, and highly deformed amphibolites (restricted to the Glenfinnan Division in this area) just east of it, so again the slide can be fairly well defined. Its suggested trace is shown on Fig. 6.1. It must be folded by the major $F3_a$ fold pair, then runs almost due north; it is shown folded by the structure north of Kinlochmoidart, since at Lochailort, it lies at the westernmost pelite band (reconnaissance and Rathbone (1980)). Note that this implies that much of what is normally assumed to be Upper Morar Psammite is actually Glenfinnan Division; also that the slide is cutting out some Glenfinnan Division lithologies (although the small pelite lenses at Kinlochmoidart may be attenuated fragments of the pelite which overlies the slide at Acharacle).

Similarly, the "Caledonian migmatite front" of section 6.4.2 can be readily located on road and coast sections. North of Loch

Shiel, it lies to the west of the Sgurr Beag Slide and is unaffected by the $F3_a$ fold pair (Fig. 6.1). It must turn NE north of Kinlochmoidart, since at Lochailort it more-or-less coincides with the Sgurr Beag Slide. It might be noted here that the Glenfinnan Division rocks around Sron na Gaoithe, like those on Mull (Chapter 7) and the Knoydart Pelite (Chapter 4) look superficially unlike typical Glenfinnan Division and "lower grade". In fact, this is due to lack of $MP3_a$ annealing after $D3_a$ deformation - although the main metamorphic distinction between the two divisions lies in the early (Precambrian?) metamorphism, the most obvious textural features (in non-migmatitic rocks) reflect the last Caledonian event. This may partly account for Brown *et al.* (1970) including the psammites around Sron na Gaoithe in the Morar Division.

From the Z-geometry of the $F3_a$ fold pair, the early isocline on Sron na Gaoithe must be folded back up around the eastern closure to run north again. There is evidence on the I.G.S. 1-inch Sheet 52 for a south-closing antiform along the axial trace of the Dig a' Bhogha fold in Morvern, suggesting that the $F2_a$ fold core corresponds to the pelite band which lies at the eastern limit of Fig. 6.2. If this is traced northwards (Figs. 6.1 & 2.1), it probably corresponds to the Lochailort Pelite, so the psammites in Fig. 6.2 and the pelites at the slide lie structurally below the Lochailort Pelite. Since there is no evidence for fold closures within either, they must correspond stratigraphically to some of the higher (eastern) units in the Glenfinnan Division.

The shape of the Dig a' Bhogha fold is itself rather unusual in that it passes north from a south-plunging antiform to a north-plunging synform (Fig. 6.6(a)). It is the north-plunging segment which is anomalous, in that regionally the Glenfinnan Division overlies the Morar Division and, from the I.G.S. map, the fold from about northing 65 southwards is an antiform. This variation of plunges through the vertical is quite common in this part of the Glenfinnan Division - e.g. Brown *et al.* (1970) describe several examples. Several possible explanations exist. It could be caused by a superimposed strain exaggerating initial slightly curving axes, or by some mechanism involving unequal advancement of points within the same axial plane - a modified simple shear model (Fig. 6.6(b), (c) - cf. Coward & Kim 1981). Rotations of more than 90° would be required in the first model, resulting from strains which would probably be incompatible with the undeformed rocks at [710672]; in the second model, the more gently plunging areas should be the most highly strained, which is the reverse of what is found. Alternatively, an interference pattern could be invoked - Type 0 of Ramsay (1967). Assuming that variable $F3_a$ plunges result from the pre-folding dips of surfaces whose strikes lay at high angles to the $F3_a$ axial planes, the form of the original fold can be inferred - it must have been almost recumbent, with a NE-dipping axial plane; the steeply plunging rocks would correspond to the core of the early fold (Fig. 6.6(d)). Since this must also fold the slide, it calls for an extra phase of folding (post-$F2_a$, pre-$F3_a$) for which there is no independent evidence; also, the most strained rocks would have come from the core of this fold, the least strained from the limbs - again the reverse of what would be expected. The model of Ramsay (1967, fig. 7-105) could produce such a structure by inhomogeneous compressive strain perpendicular to the axial plane (Fig. 6.6(e)). This has the advantage of producing maximum strain in the steeply plunging rocks, as observed (section 6.2.1); it should be noted that the platy/non-platy criterion in these rocks probably depends mainly on the finite value of $\underline{x}$, and so, for example, extensions parallel to the hinge at A on Fig. 6.6(e) would not be recorded. All these models involve varying amounts of sub-horizontal, N-S extension, which would probably be unrealistic if applied over the whole orogenic belt. However, particularly with the fourth model, compensating strains could occur to the north or south. Two other possibilities can be mentioned, which give the platy, reclined zone more significance. If it was a post-folding simple shear zone, it could have juxtaposed two entirely separate folds from different levels (movement would be vertical, since axes become vertical at high strains). However, this might be expected to result in displacements of major structural and stratigraphic features (e.g. the Sgurr Beag Slide), or a contrast in $D3_a$ structure and metamorphism across the zone. Alternatively, the two folds could have originally been steeply plunging, and have somehow been twisted about their common limb, giving rise to the platy zone. No obvious mechanism exists for such a process, and the fact that the Dig a' Bhogha fold apparently passes south into an antiform may cause problems. There is clearly no simple solution to the problem, but the most likely explanation is that it is due to some inherent feature of the $D3_a$ deformation: of the models proposed, that of Fig. 6.6(e) poses the less serious objections.

Regional structural correlations are discussed in detail in Chapter 14; however, from the descriptions in this chapter, it is likely that $D1_a$, $D2_a$ and $D3_a$ structures correspond to the D2, D3 and D4 structures of Mallaig, and that the S-S1 bedding schistosity of Mallaig was not recognised here due to the intensity of later deformation.

CHAPTER 7

Mull

7.1 Introduction

Moine metasediments on Mull are restricted to several small coastal inliers and to one area of about 17 square kilometres in the Ross of Mull. There, they are bounded to the west by the late Caledonian Ross of Mull granite (c. 420Ma - Beckinsdale & Obradovitch 1973), and to the northeast are in fault contact with Tertiary lavas, the latter overlying a thin Mezozoic succession (Bailey & Anderson 1925). The area lies well to the west of the Glenfinnan Division proper, and may be close to the continuation of the Moine Thrust - at least, Lewisian and possible Torridonian rocks occur on Iona, west of the granite (Figs. 2.2, 7.1(a)). A.L. Harris (verb. comm.) had noted a strong similarity between the striped and pelitic rocks of the Ross, and the Glenfinnan Division, and between their eastern boundary, south of Scoor House [419191], and the Sgurr Beag Slide. This was confirmed by Rathbone (1980), who studied the slide zone. Good exposure is restricted to the south coast, where broad wave-cut platforms make detailed structural studies possible.

Two contrasting groups of lithologies are present, separated by a major zone of high strain, the Scoor Slide (section 7.4). To the east a thick succession of trough cross-bedded, feldspathic psammites (Plate 7.1(a)) closely resembles the Upper Morar Psammite (cf. Chapter 2). This passes eastwards into garnetiferous pelites and further psammites, possibly comparable to the Morar (Striped and Pelitic) Schist and Lower Morar Psammite. Abundant sedimentary structures indicate consistent westward younging, at least as far as the slide zone. Current directions, indicated by the trough axes, are northwards and upwards, in rocks which are sub-vertical and trend N-S. The succession to the west of the slide commences with 80m of laminated quartzites, with bands of pelitic gneiss up to 1m thick appearing towards the west. This is followed by over 300m of pelitic gneiss, containing 10cm thick lenses of garnetiferous semipelite. The remainder of the section is occupied by a striped series with quartzose psammites and muscovite-rich pelites in varying proportions. These rocks have some affinities with a turbidite sequence (P.J. Brenchley, verb. comm.). Sedimentary structures are generally absent, but in a few localities, possible graded bedding indicates that the rocks are downward facing in the major $F4_r$ structure, and so would young towards the slide (section 7.2.4). At one locality, [418186], small-scale cross-bedding was seen which was upward facing in $S4_r$. This lay in an area of $F1_r$ and $F2_r$ folding, and so may not be regionally applicable. However, somewhat further west, Rathbone (1980) recorded convolute lamination which agreed with the cross-beds. Rare kyanite-bearing pelites occur (e.g. [378189]), as do dykes of garnetiferous amphibolite (Plate 7.2(a)), which predate the peak of metamorphism and are identical to those in Salen Bay (Chapter 6). Calc-silicate lenses occur in the more psammitic rocks. In virtually all aspects these rocks are typical of the Glenfinnan Division; psammites and semipelites are somewhat less coarse than usual, but as at Acharacle (Chapter 6), this is a secondary feature reflecting lack of post-$D3_r$ and $D4_r$ annealing. At the northeastern margin of the granite, Morar Division psammites reappear northwest of the Glenfinnan Division, some outside the intrusion, others as large xenoliths or pendants preserving a ghost stratigraphy (Fig. 7.1(a)). A similar strain pattern to that a Scoor is seen in southward traverses (A.L. Harris, verb. comm.), and the first Glenfinnan Division lithology is a quartzite similar to that at Scoor Bay (Bailey & Anderson 1925, p. 25). This fits in with structural evidence (section 7.2) which suggests that the Glenfinnan Division rocks occupy a south-plunging synform. Apart from the granite and amphibolites, igneous rocks include foliated microdiorites (which predate the granite), alkali basalt and dolerite sheets of Permo-Carboniferous age (c. 275Ma, Beckinsdale & Obradovitch 1973), and Tertiary dolerite dykes.

7.2 Structure

In any given place, it is usually possible to demonstrate the presence of three main phases of deformation, of which the third gives rise to the dominant mesoscopic structures. Both the major Assapol Synform and the Scoor Slide are of local F3 age on this basis, but it seems likely that the synform folds the slide on a large scale, so they will be called $F4_r$ and $F3_r$ respectively. This is discussed further in section 7.2.3. $F2_r$ structures are mainly seen in interference patterns, although $S2_r$ is the dominant foliation in the pelitic gneiss of Scoor Bay; $F1_r$ structures are restricted to a bedding foliation and rare isoclinal folds.

7.2.1 $F1_r$

Throughout most of the area, the main evidence for $F1_r$ is provided by intrafolial folds in the $S2_r$ fabric, preserved where the later $F4_r$ structures are relatively open. Where $F4_r$ structures are tight, it is not generally possible to distinguish between $F1_r$ and $F2_r$ minor folds. Examples are seen in the striped rocks at Uisken (Plate 7.3(c), [395187]), where the $S2_r$ fabric, in the core of a 20m-wavelength $F4_r$ fold, is a crenulation cleavage. In the pelitic gneiss west of Scoor Bay (loc. (g) on Fig. 7.1(b)), where $F4_r$ structures are virtually absent, rare isoclinal folds of bedding occur, overprinted by the penetrative $S2_r$ gneissosity (Plate 7.2(b)). At locality (h), isoclinal folds of bedding occur, again in an area where $F4_r$ deformation is weak. These are overprinted by an $S2_r$ foliation, which locally is clearly a crenulation cleavage (Fig. 7.2(a)). The folds have northerly plunges ranging from zero to 60°; both $F1_r$ and $F2_r$ minor structures indicate a major antiform to the west. Some of the $F1_r$ fold cores seem to be eliminated by sliding.

7.2.2 $F2_r$

This generation of structures is widely represented. The foliation which is crenulated in the ubiquitous $F4_r$ folds is of this age, and locally (in the area west of Scoor), it is dominant. Occasional intrafolial folds are seen (e.g. in Plate 7.3(c)) but the foliation is usually a penetrative, differentiated schistosity (e.g. Plate 7.3(d)). The penetrative gneissose fabric in the migmatitic pelites at Scoor is of this age (Plates 7.2(c), 7.3 (e)).

The lits appear to have segregated in S2, and pass into smaller quartz lenticles in the semipelites; even in protected areas (e.g. the $F1_r$ fold core in Plate 7.2(c)), transposition is not in evidence. A quartz vein with minor feldspar is folded about S2, but such veins occur in the lithologically-similar (but garnet grade) garnetiferous pelite at Mallaig, and need not be related to migmatisation. Larger folds of this age are quite common in the pelitic gneiss (e.g. Plate 7.2(d), loc. (g) on Fig. 7.1(b)). Plunge (on steep, N-S $F4_r$ limbs) varies from 0/020 to 50/040; vergence invariably indicates an antiform to the west. Minor folds with this vergence and a penetrative axial planar fabric are widespread throughout the eastern limb of the $F4_r$ Assapol Synform (Fig. 7.2(b)); they are probably present on the western limb, but cannot readily be distinguished from tight $F4_r$ structures with the same vergence. In the core of the Assapol Synform [398188], an early penetrative fabric is seen, whose vergence is consistent with these $F2_r$ structures (Fig. 7.2(d)).

7.2.3 $F3_r$

The Scoor Slide occurs at the boundary between the cross-bedded, arkosic psammites and Glenfinnan Division quartzites, east of Scoor Bay (Fig. 7.1(b)). It shows the strain pattern typical of the Sgurr Beag Slide (Rathbone & Harris 1979, Rathbone 1980) - briefly, planar discordances are eliminated, the rocks become very fissile and platy, folds become reclined and are then eliminated towards the boundary. The strain pattern is discussed in section 7.4; here, only textural evidence relating to its age will be presented. In the Morar Division psammites to the east of the slide zone, no strong fabrics of any identifiable age are seen. However, a coarse, polygonal texture, similar to that attributed to MP2 at Mallaig, is broken down in the slide zone (Rathbone 1980). In the Glenfinnan Division, the pelitic gneiss carries a strong $S2_r$ fabric, associated with the migmatisation and peak of metamorphism, which is cut by $MP2_r$ garnets. This is progressively overprinted in the easternmost 50m by a platy foliation, sub-parallel to bedding, and in the last 10m or so, the rocks are very platy and fissile, as are pelites within the quartzite unit. The migmatitic lits are highly attenuated, with ribbons of strained quartz wrapping feldspar porphyroclasts, and garnets are strongly augened (Plate 7.3(f)). The slide therefore appears to be local F3 in age ($F3_r$). The relatively low metamorphic grade (section 7.3.2) makes it unlikely that the slide represents a local intensification of $D2_r$ deformation, since time is required for the Glenfinnan Division rocks to cool (probably by 100-200°C) between the $MS2_r$-$MP2_r$ metamorphism and sliding.

The relationship between the slide fabric and $F4_r$ structures is not so clearly seen in this section. Throughout the area, from locs. (a) to (g) on Fig. 7.1(b), $F4_r$ crenulations are weak or absent, and no $F4_r$ minor folds occur. However, since the Morar Division reappears on the western limb of the $F4_r$ Assapol Synform (section 7.2.4), and is highly deformed at its boundary, the major fold must post-date the slide. Any possibility that the slide was synchronous with the Assapol Synform is ruled out by the extremely large vertical displacements which would be implied (section 7.4.2). Open crenulations of the slide fabric occur at locs. (e) and (f), dipping eastward at 60-70°. These may be weak $F4_r$, but it is possible that they are relatively tight examples of the late crenulation (section 7.2.5).

7.2.4 $F4_r$

This generation gave rise to the most prominent folds on all scales. A major tight synform in the Glenfinnan Division, the Assapol Synform, is defined by medium-scale $F4_r$ folds ranging from a few metres to a few tens of metres (Plate 7.3(a), (b)), and confirmed by the symmetrical distribution of lithologies (Bailey & Anderson 1925, chapter IV). These folds are upright, with steeply ESE-dipping axial planes and moderately SSW-plunging axes. West of "A" on Fig. 7.1(a), they show consistent Z-profiles, while to the east, S-profiles are seen. The main synformal closure lies in the bay of Slochd Mhi Chriscain [398188]. $F4_r$ plunges in this region tend to be shallow or even horizontal; they steepen on the limbs. Orientation data for the central part of the synform [392188 to 403188] are shown in Fig. 7.3. This indicates that the fold axis plunges at about 25/205, on an axial plane dipping at about 80/120. The plunge of minor folds varies within the $F4_r$ axial plane, presumably due either to varying initial orientation of bedding, or varying superimposed strain. The association of steeper plunges with limbs favours the latter explanation. The $S4_r$ fabric is usually a tight crenulation cleavage, but ranges from a penetrative schistosity on some limbs to an open crenulation in some cores (Plate 7.3(c), (d), (g)). Micas in crenulation cores are usually recrystallised. Interference patterns with $F2_r$ are common on the eastern limb of the synform (Fig. 7.2(c)); the general absence of $F3_r/F4_r$ interference is due to the restricted nature of the $S3_r$ fabric.

7.2.5 Later structures

A widespread late crenulation overprints all structures up to $F4_r$. It usually dips at moderate angles to the SE (Fig. 7.3), but locally is almost horizontal; it is approximately coaxial with $F4_r$. The crenulations are open (interlimb angle $>90°$), and micas are bent or broken on crests (Plate 7.3(h)).

Some open folding on upright, E-W axial planes occurs in the area southeast of Scoor and causes variation in the strike there. Kink bands and brittle folds are associated with this deformation.

7.3 Metamorphism

7.3.1 Pre-slide metamorphism

Due to later events, it is not possible to determine metamorphic grade during $F1_r$. In the Glenfinnan Division, the metamorphic peak seems to have been approximately synchronous with $D2_r$. In the migmatitic pelites, $S2_r$ is a coarse, gneissic fabric with segregation of quartzofeldspathic lits; large, essentially unzoned garnets cut the foliation and the leucosome/melanosome boundary. These are reworked at lower grade by $D3_r$ structures in the slide zone. In other lithologies, garnet is augened by $S4_r$ (e.g. (58134)); kyanite in (59561) is also early. It is likely that kyanite, garnet, the gneissic fabric and most of the coarse mica in the Glenfinnan Division date from the one, pre-sliding, $MS2_r$-$MP2_r$ metamorphic episode. The presence of kyanite and the migmatitic textures suggest amphibolite facies conditions; P,T estimates of c. 600°C at 6.5kb are comparable with the Glenfinnan Division elsewhere (section 15.2.3(v)).

Grade in the Morar Division can only be roughly estimated - biotite is certainly present, and small garnets in semipelitic intercalations. In general appearance they are comparable to the garnet grade rocks of Morar and western Ardnamurchan.

7.3.2 Syn- to post-slide metamorphism

This metamorphism was relatively uniform throughout the area; no new garnet grew. $S3_r$ textures are restricted to the slide zone, where a strong muscovite-biotite foliation which augens garnets is developed. Micas in rare intrafolial folds are recrystallised, but quartz and feldspar in lits are highly strained. The ½-1mm polygonal texture in psammites is reduced to a much smaller grain size through the development of sub-grains. Micas in $F4_r$ crenulations are usually recrystallised, particularly when the crenulation is tight (Plate 7.3(g)). In some open crenulations, the micas are bent through about 20°, then recovered (Plate 7.3(d)). This probably indicates biotite grade for $MS3_r$-$MS4_r$. Temperatures must have fallen by the time of the late crenulation, since micas are always bent or broken, never recrystallised (Plate 7.3(h)).

The Ross of Mull granite gave rise to andalusite in its aureole (e.g. replacing the kyanite in 59561) and sillimanite in xenoliths, indicating a typical high temperature, low pressure environment. From Fig. 7.1, it must be later than the Assapol Synform, and $F4_r$ structures are annealed in the aureole. It is probably much later than all the structures discussed above, since it seems to be about 20Ma younger than the foliated microdiorites (c. 440Ma, van Breemen et al. 1974), which themselves cut these structures.

7.4 The Scoor Slide

7.4.1 Description

The Scoor Slide shows many features typical of the Sgurr Beag Slide elsewhere in the Highlands (Rathbone & Harris 1979, Rathbone 1980). It forms the boundary between migmatised, Glenfinnan Division-like, and unmigmatised, Morar Division-like rocks, and lies at the centre of a zone of high strain, shown by attenuation of bedding and sedimentary structures, development of a strong platy fabric, rotation of linear structures into a reclined position, and augening of migmatitic and high grade metamorphic textures. The study was based on an east to west traverse along the coast east of Scoor Bay (Fig. 7.1(b)). Bedding is sub-vertical and strikes north-south.

Between 400m and 200m from the slide, the Morar Division consists of thick (2m), trough cross-bedded arkosic psammites, consistently younging to the west. These are separated by very thin pelitic partings. The troughs themselves are defined by micaceous partings which weather in (Plate 7.1(a)). Trough axes indicate northward and upward transport of sediment in this area. Undeformed beds alternate with more highly strained zones a few metres thick, within which cross-beds may be eliminated, and a strong fabric is developed in the more micaceous rocks. The last unstrained psammites occur at loc. (a) on Fig. 7.1(b), approximately 200m from the slide. In the next 50m, to loc. (b), all psammites are noticeably deformed, but strain remains heterogeneous, i.e. zones in which troughs are clearly recognisable (Plate 7.1(b)) alternate with more platy zones. Ptygmatically folded quartz veins become common (usually indicating >50% shortening), beds are always <1m thick, and maximum cross-bed angle is 10-15°. The platy zones have a notably lenticular appearance on horizontal surfaces, with inter-lens boundaries weathering in; they look like highly deformed troughs, and measurements bear out this suggestion (section 7.4.2). From 150m to 100m, recognisable troughs are more strongly deformed (Plate 7.1(c)), beds are thinner (<30cm), and maximum cross-bed angles are down to 5-7°; the proportion of the section which is "platy" increases westwards. In the following 50m, no clear troughs (with internal laminations) are preserved; this lithology is now only represented by the lenticular platy rock. Bed thicknesses are now 10cm or so, and ptygmatic veins indicate large degrees of shortening (>80%). The platy rocks themselves become more attenuated - the least deformed rocks between (c) and (d) on Fig. 7.1(b) are comparable to the most deformed rocks between (b) and (c). In the final 50m, the rocks are uniformly platy with no discordances (Plate 7.1(d)).

The line of the slide can only be located as a lithological boundary; there are no distinctive structural features at it, but the transition from arkosic psammite to quartzite can be defined within 2-3m on the ground. The quartzite is perfectly planar, and banded on a centimetre scale, with planar internal laminations 1-2mm apart. Within a few metres, a highly deformed migmatitic pelite band a few decimetres thick appears (58141). To the west, the rocks rapidly become less platy, and by loc. (e) quartzite beds are 10m thick; every few metres a pelite (still platy) occurs, containing 1cm X 50cm lenses of quartzite. By loc. (f), at 50 metres, quartzites are 10-20cm thick, with internal laminations a few centimetres apart, and thickness of pelite bands has increased to a few metres, although they are still platy. Over about 20m, there is a transition (by increasing frequency of the pelite bands) to pelitic gneiss. By this point, garnets and feldspar augen are obvious in hand specimen, although a strong bedding-parallel fabric is still present. The pelitic gneiss unit consists of homogeneous migmatitic pelite containing 10cm stripes of semipelitic schist. The lower 10 metres of this are quite platy, but then it rapidly becomes very coarsely migmatitic and within 20 metres, an $S2_r$ gneissic fabric, oblique to bedding, is preserved.

There is therefore a clear strain gradient towards the slide from both east and west. In the Morar Division, strain is highly heterogeneous, in apparently uniform lithologies, except for the last 50m. Presumably, in the absence of any weak horizons, initial slight variations in deformation have been amplified by some process of strain softening. The Glenfinnan Division side appears to have suffered less deformation than the Morar Division side - the broad platy zone is not present, and strain in the quartzites decreases rapidly after the first few metres. This might be reasonable if the slide was comparable to a thrust or mylonite zone, with the Glenfinnan Division on top (i.e. most deformation occurring in the footwall). Alternatively, the pelitic interbeds may have accommodated most of the strain, possibly with slip along the foliation, leaving relatively little ductile strain to register in the quartzites. There is some evidence to favour this hypothesis - even when quartzites are 1m or more thick, intervening pelites are still platy; the fact that the slide deformation dies out in the lower 20m of the main pelite suggests that there may effectively have been décollement at this boundary (cf. Kinlochourn, section 5.2.1, where the slide fabric dies out rapidly upwards into a pelite).

7.4.2 Strain estimates

Some attempts were made to estimate the strains associated with the Scoor Slide. Data are rather limited, and mostly two-dimensional, so a number of geometric assumptions have to be made. The method itself is prone to large errors (perhaps +100%,-50%), and the estimates do not include any displacement on discrete shear surfaces in platy zones. In addition, the assumptions made in converting measured shortening to shear strain mean that only a minimum estimate can be obtained. Even if the estimates are only in the right order of magnitude, they will still be useful.

Initial studies were based on the shape of troughs in the cross-bedded Morar Division psammites. These show a clear increase in deformation towards the slide, with flattening of troughs and reduction in angle of internal cross-laminations (Plate 7.1(a) - (c)). Time did not permit a full statistical study, so at each locality, beds were selected for measurement, which appeared to lie near the middle of the range of geometries observed. Beds showing planar cross-bedding were excluded. Measurements of length/width ratios of troughs were made on sub-horizontal joints, perpendicular to bedding (Fig. 7.4), and the means of the ratios were compared in deformed and undeformed rocks. This indicates attenuation by a factor of two for the cross-beds in Plate 7.1(b), and five for those in Plate 7.1(c). The mean length of the troughs was also compared, with the result that there was no significant difference between strained and unstrained rocks (data in Fig. 7.4), values ranging around 40cm (the t-test is not strictly valid for bed B, since the standard deviations are significantly different). This indicates that the deformation involved little or no horizontal extension within the bedding. If approximately constant volume was preserved, there must have been considerable vertical extension (there is no evidence for pressure solution, such as accumulation of mica or opaques, and since the rocks had already been at garnet grade it is unlikely that any dewatering process was involved). This could not be assessed quantitatively; however, in undeformed rocks, trough forms are seen in vertical, east-west faces, with similar shapes to those on horizontal surfaces (i.e. the trough axis must plunge at a moderate angle to the south). In deformed rocks, lamination is quite planar in vertical faces, even when troughs are seen on horizontal surfaces. This indicates that the two-dimensional strain observed in vertical faces is higher, i.e. that there is strong vertical extension, with the trough axis being rotated into a near vertical position. Also, the probable deformed troughs in the lenticular platy rocks are blade-shaped in three dimensions - i.e. a few millimetres across, half a metre wide and several metres high. It is noteworthy that, even in these highly strained rocks, the length of the troughs is little changed, i.e. there is still very little horizontal extension. It can be assumed that bedding in the most highly deformed rocks (with > 90% shortening) effectively coincides with the xy-plane of the strain ellipsoid. A likely implication of this would be that the slide deformation was basically a plane strain, with xz-plane sub-vertical, trending east-west. The data would fit well with a simple shear zone with up-dip translation.

The strain estimates will be approached in two parts - first, based on the above observations, the strain in a sub-horizontal section can be represented by t/t', i.e. the factor by which an element has been thinned, assuming no horizontal extension. This will then be converted into shear strain by making some further assumptions.

The three sets of deformed cross-beds discussed above (Fig. 7.4) represent the least strained rocks at their localities. These are plotted in Fig. 7.5 as minimum values. The maximum strain in any area occurs in the platy zones - a number of estimates are shown, based on lenticular troughs (assuming an original length of 40cm, and width of 5cm), and also on maximum thickness in a group of about twenty beds (in undeformed rocks, five to ten of them would be one to two metres thick). Some estimates were also made based on maximum cross-lamination angle (30-35° in undeformed rocks). All these methods were tried on the three beds measured in detail, and agreed within 20-25%. The outlying platy zones (>200m) show shortening of 5 to 10 times, which agrees with the fact that by bed D with $t/t' = 7$, cross-lamination is barely recognisable. Shortening indicated by ptygmatic veins was recorded in places; it was taken to give average values of strain in a given area, although if the vein was emplaced during the deformation, it would not have recorded all the strain. Four possible plots of t/t' are given in Fig. 7.5. Curve (i) may represent the maximum strain locally - the highest values are based on areas of a few metres in which no bed is thicker than one centimetre. Curve (iv) may represent the minimum local strain, i.e. the least deformed beds in the last 50m look subjectively as platy as the most deformed rocks at 80-100m. Curve (ii) is drawn midway between, and may be an approximation to the mean strain over 10-20m segments. Curve (iii) is a modification of this, which takes into account the relative uniformity of the platiness in the last 50m, possibly implying a flat-topped profile. This is probably the best choice. Assuming plane strain, and no internal duplication, curve (iii) can be integrated to give the pre-deformation thickness of this section of Upper Psammite. The result of 4km, added to the 1.5km between here and the probable Morar Pelite, is reasonable in the context of Moine stratigraphy (e.g. a minimum of 6km is present at Arisaig).

If this strain is due to a simple shear zone with the orientation suggested above, the shear strain can be calculated fairly readily. Taking the model of a planar bed intersecting a shear zone boundary at an angle α , Fig. 7.6 shows how the measured thickness is changed. Fig. 7.6(a) shows the case of measurement perpendicular to the shear zone (relevant to wave-cut platforms in the platy zones), and Fig. 7.6(b) the case of measurement orthogonal to bedding (relevant to thicker beds with joints). For low values of α, the results are virtually identical. Since the troughs are highly flattened in bedding initially, and interfaces in the critical area for thickness measurement are sub-parallel to bedding, this model is probably adequate for them (particularly at higher strains). Since bedding in the platy rocks, the xy-plane, and the movement plane in the shear zone will all be sub-parallel at high strains, the geometrical relationship of platy zones to less deformed or undeformed rocks provides some indication of α . On horizontal surfaces, platy zones are parallel to bedding over tens of metres. Although vertical exposures are usually limited to a few metres, deviations of more than about 10° would have been easily recognised, and 20° or more would have been obvious. If the structure is analogous to a thrust (see below) angles of a few degrees would be likely. The ratio of γ to t/t' for $\alpha = 5°$ and $10°$ is shown in Fig. 7.6, and plotted in Fig. 7.7. This was used to calculate displacement on the Morar Division half of the shear zone, by $s = \int \gamma dh$, where s is displacement, γ shear strain and h horizontal distance on the ground. For curve (iii), $\alpha = 10°$, displacement is 21km, and for $\alpha = 5°$, 42km. Assuming $\alpha = 20°$ or 2.5° would give about 10km and 80km respectively. Clearly only a minimum estimate can be obtained, since α = zero gives s = infinity. 40km might be taken as a reasonable estimate. If the Glenfinnan Division had an identical strain profile, an overall displacement of about 80km might be indicated. Based on quartzite bed thicknesses, high strains comparable to the last 50m of the Morar Division are restricted to the first 10m or so, and there is then a rapid decrease to t/t' of about 10; very rough calculations of displacement give results of 5 to 10km. However, movement has probably been concentrated at the pelite interbeds, so that the quartzites give a gross underestimate of the strain (section 7.4.1). Bearing in mind the possibility that some movement has taken place at platy horizons in the Morar Division, doubling the half displacement is probably reasonable.

To summarise, the strain pattern is consistent with the slide being a simple shear zone with about 80km displacement (assuming $\alpha = 5°$), in an up-dip direction (in the present orientation). Clearly such displacements could only be accommodated in a shallow-dipping zone equivalent to a thrust. Since the present steep dip is due to these outcrops lying on the limb of the later Assapol Synform, this would be acceptable. A generally east-to-west movement (pre-folding) would be implied. In strike, movement direction and displacement, the Sgurr Beag Slide in general invites comparison with the Moine thrust zone (for which Elliot & Johnson (1980) now suggest 77km displacement). If the Moine Thrust does lie east of Iona, the Sgurr Beag Slide must be earlier, although its relations to the Moine mylonites would be less easy to define. Some independent support for the suggested displacements is found by treating the Ross of Mull Glenfinnan Division as a klippe. The present outcrop distance from the western limit of the Glenfinnan Division on Mull to the Glenfinnan Division in Morvern (I.G.S. 10 mile map) is about 30km, not allowing for any shortening across the Morar Antiform and Assapol Synform. This would be a minimum estimate, since the Ross of Mull rocks are still distinct in lithology and metamorphic grade from the Morar Division just west of the Slide at Acharacle (Chapter 6). P,T estimates indicate that the Morar and Glenfinnan divisions come from similar structural levels (section 15.3.2), severely limiting any vertical displacement on the slide. Thus the important point is not the precise displacement, but the fact that it is several tens of kilometres rather than one or two kilometres. Only the latter would allow an initially steeply-dipping plane, so any possibility that the slide is related to high strain on the limbs of the Assapol Synform is ruled out.

7.5 Conclusions

The situation of the Glenfinnan Division on Mull is comparable to that on the mainland. Kyanite grade metamorphism, and migmatisation of certain horizons occurred at the climax of two phases of deformation. These rocks were then brought into contact with garnet grade Morar Division across the Scoor Slide. The tight, upright Assapol Synform was then formed, in the core of which the Glenfinnan Division is preserved as an outlier. Both these events took place at biotite grade. Displacement on the slide was several tens of kilometres (possibly about 80km), probably with westward overthrusting of the Glenfinnan Division. The Scoor Slide is directly comparable with the Salen Slide at Ardnamurchan, with which it probably connects across the southern continuation of the Morar Antiform, which itself is probably complementary to the Assapol Synform (cf. Chapters 3, 4, 14).

In the calculations of section 7.4.2, it was assumed that bedding intersected the slide zone at a small positive angle. If it had a small negative angle (e.g. $\alpha = 170°$), the qualitative results would be unchanged, since the curves relating to α converge on those relating to $(180° - \alpha)$ at moderate strains (e.g. by t/t' = 10 they are within 20% of one another (Fig. 7.7)). In fact, it would result in higher values for the displacement. However, in that case, some form of turnover would be expected at quite low strains, where beds would apparently be thickened. Possibly these would be observed as folds - in this context, it may be significant that no folds are seen in the psammites here (cf. other areas - e.g. sections 5.2.1, 8.2.2). On unfolding the $F4_r$ structures, the slide would cut downwards into bedding to the west. However, since the rocks were already folded, it could still be cutting upwards in relation to some external datum, e.g. sea level.

CHAPTER 8

Sguman Coinntich

8.1 Introduction

The peak of Sguman Coinntich lies in southern Ross-shire, some 5km east of the head of Loch Long (Figs. 2.1, 8.1). The main ridge is formed by an elliptical outcrop of migmatitic pelite of Glenfinnan Division aspect, which is surrounded by Morar Division psammites; it lies 12km east of the Moine Thrust and 6km east of the northward continuation of the Glenelg Lewisian. The relevant I.G.S. 1-inch sheet is not published, but the area formed part of that studied by Clifford (1958a). He interpreted it as a synformal outlier on a NE-SW axial plane, plunging inward at both ends due to later refolding. More recent work (A.L. Harris, P.A. Rathbone, verb. comm.) suggested that it might be an outlier of Glenfinnan Division rocks, resting on the Sgurr Beag Slide, possibly comparable with the larger Beinn Dronaig structure 10km to the northeast. However, in view of the conclusion reached below that the pelite lies in an antiformal core, the regional structure cannot be so simple.

Sguman Coinntich was studied on the basis of detailed traverses in critical, well-exposed areas, and only reconnaissance mapping of boundaries elsewhere. Efforts were concentrated to the south of the main fault (Fig. 8.2), along the north-western pelite boundary, and eastwards from [978293]. North of the last, the exposure consists of slabs parallel to the main $S3_s$ cleavage, many of which have slipped downhill, while detailed structures on the summit are obscured by frost-shattering.

Two main lithological units can be recognised - psammite (locally migmatitic) and highly migmatitic pelite.

Migmatitic Pelite This is a coarse, homogeneous, *lit-par-lit* migmatitic pelite, of typical Glenfinnan Division type (Plate 8.1(a), (g)). Locally it is interbanded with semipelites on the scale of a few centimetres (Plate 8.1(b)). *Lits* are 1-5cm thick, have a few centimetres grain size, and may range in length from several decimetres to several metres (Plate 8.1(a), (g)); the feldspar is almost invariably oligoclase, and usually forms about 50% of the lit. The matrix is coarsely schistose and muscovite-rich, and contains abundant pink garnet crystals up to 1cm in diameter, as well as small quartzofeldspathic lenticles (e.g. 56233). Small lenses of garnetiferous amphibolite are present (e.g. [966287], [977303]), consisting largely of hornblende and garnet (few millimetres grain size), with accessory epidote, sphene and ilmenite, and quartz-plagioclase (An_{45}) veins. Such lenses are generally considered to be diagnostic of the Glenfinnan Division (e.g. Powell 1974, Johnstone et al. 1969).

Psammite This covers a wide range of types, including quartzite, feldspathic psammite and micaceous psammite. A distinctive quartz-biotite rock also occurs, forming a large area in the vicinity of [987293] and extending to NE and SW; in the latter area [973282], it interdigitates with the other psammites. Small hornblendic lenses (striped amphibolite and hornblende gneiss) occur within the psammite, and are almost certainly Lewisian inliers, possibly resting on subsidiary slides (e.g. at [98452923], [98592942]). These define several concentric shells, also marked by slight variations in psammite composition, which surround the Sguman Coinntich and Beinn Dronaig pelites (A.L. Harris, verb. comm.); mapping here was not sufficiently detailed to delineate them. The massive, often feldspathic nature of the psammites, contrasting with the typical striped pelite/quartzite Glenfinnan Division, and their apparent continuity with psammites above the Glenelg Lewisian, make it likely that these are Morar Division psammites. Migmatisation, where present, is late, i.e. usually post-sliding, and often cross-cutting (section 8.4). Sedimentary structures are absent, due to the high deformation in the area studied.

The boundary between the psammite and pelite is extremely sharp (Plate 8.2(a)), and lies at the centre of a zone of high deformation, which probably represents the Sgurr Beag Slide (section 8.2.2).

Dips are generally to the south-east, parallel to the limbs of the major tight folds, and a SSE-plunging lineation is common.

Two major E-W faults occur (Fig. 8.2); the southern one is more clearly defined, and is exposed in the Allt a' Glas-choire at [98452923]. Assuming that it is planar, its orientation can be gauged from the topographic effect in the region of [966293], where it appears to dip south at about 45/185. The trace of the northern fault is less well defined, but appears to be convex northwards, indicating a moderate southerly dip. A number of minor N-S faults occur, and these are essentially vertical; the largest is exposed in the stream at [97022870] and displaces the pelite boundary by about 20 metres (apparent sinistral displacement; not shown on Fig. 8.2).

8.2 Structure

The major structure is the $F3_s$ Sguman Coinntich fold, and the dominant minor structures are parasitic on it; it is most convenient to describe the structures in reverse order, i.e. $F3_s$, $F2_s$ (slide), $F1_s$.

8.2.1 $F3_s$

In the core of the Sguman Coinntich fold, the Glenfinnan Division pelite forms an elliptical outcrop, with long axis trending NE-SW (Fig. 8.2). Subareas within the fold can be defined as follows: the NE and SW core zones, north and south respectively of the two major E-W faults; the NW and SE limbs, and the central core zone, these lying between the major faults. The NW and SE limbs both dip at moderate angles (30-50°) to the SE or SSE; the NE and SW cores dip steeply to the NE or SW. At the present level of exposure, the structure has the form of a flattened cylinder, dipping and plunging to the SE; it must close upwards or downwards to take the form of a flattened test-tube or sheath fold (Quinquis et al. 1978). The main question to be decided is whether it is an antiform or a synform.

Both major and minor structures indicate that it is an antiform. The evidence is clearest in the NW and SE limbs, since at both ends the fold is sideways-closing. Comparing the two limbs, the NW clearly has steeper dips of bedding and migmatitic banding - this is confirmed by the stereogram (Fig. 8.3(a)). Although there are relatively few readings, the NW limb ranges from 45 to 65°, while the SE limb ranges from 25 to 45°. The spread is partly due to later folding (section 8.2.4) - e.g. the steep dips at [983296] lie on a limb of a NW-SE trending open fold. Comparing two points opposite one another, and with similar strike, eliminates this effect, so that, for example, the NW limb dips at 50° at [969295] and the SE limb dips at 30° at [974290]. Poles to the axial planar cleavage (see below) lie intermediate in dip between the two limbs (Fig. 8.3(b)).

Vergence of minor structures agrees with this conclusion. Throughout the area, the dominant minor structures are tight-to-open folds with axial planes dipping at 20-40° to the SE and having variable plunge (Plate 8.1(a), (b), (e), (f), (g)). In the south-west, they are almost reclined (usually pitching 70-80°S on their axial planes), and display a systematic change in vergence related to the major closure, changing south-eastwards from S-profile through M and W to Z (Fig. 8.2). North of the major fault, the pitch of the fold axes decreases, and they become notably curvilinear, changing from reclined, through southward pitches, to horizontal. Around [970296], the folds pass through the horizontal, and in the north-east, they steepen again through northward pitches to reclined, so that folds on the NW limb now have Z-profiles on the ground. Clearly, if the major fold axis follows the minor axes, the structure must be an antiform. In addition, along the NW limb, there are numerous examples of $F3_s$ folds with Z-profiles looking north (Plate 8.1(c)) and gentle plunges, and the $S3_s$ crenulation cleavage is consistently shallower than bedding. On the SE limb, suitable (vertical, NW-SE) exposures do not occur, but around [981293], $F3_s$ folds are relatively open, and exposures are parallel to an enveloping surface indicating overall bedding. This dips southward, while the cleavage dips SE - one example of this is shown in Fig. 8.3(c), and clearly indicates that the exposure is on the upper limb of a south-plunging antiform.

Several other lines of evidence could be mentioned. For example, the main E-W fault must be normal from the eastward displacement of the northern side, given the general south-easterly dips (note that the thin sliver of pelite from [962293] to [968293] consists of small crags surrounded by peat, and so may lie in a broader fault zone, i.e. the NW boundary on the northern side may be displaced eastward by up to 600m). The pelite outcrop is narrower on the downthrown side, suggesting that the fold closes upwards, i.e. is an antiform. A similar situation obtains at the northern fault. Standing anywhere around [970296], and looking ESE across the hillside, it is clear that most exposures larger than a few tens of metres in the [978294] area dip southwards at about the same angle as the hillside, before swinging to a NE strike near the contact; thus the structure there is a south-plunging antiform. A moderately strong LS fabric is present in many psammites, reworking the earlier platy fabric, and appears to be axial planar to $F3_s$ folds (Fig. 8.3). In the SW core zone, the L component pitches about 80° south in the axial plane. Where bedding/cleavage intersections are seen in the nearly-reclined folds of this area, they have slightly lower pitch on $S3_s$, i.e. lie further SW (the angle is too small to be seen in the synoptic stereograms). If the extension lineation is parallel to the long axis of the "test-tube", and fold axes are rotating towards it, then this is the pattern that would be expected from an antiformal closure.

Small-scale $F3_s$ structures, as has been mentioned, vary in style and plunge. Their orientation ranges from reclined to horizontal, on fairly uniform SE-dipping axial planes (Fig. 8.3). They are usually fairly tight, with a strong axial planar crenulation cleavage, and deform both migmatitic banding (in the pelite) and bedding (Plate 8.1(a), (b)). In the SW area, the core of the major fold contains several 100m-wavelength $F3_s$ folds, so it consists of large areas with M and W folds, separated by relatively narrow limbs with S and Z folds. There is a tendency for these limb zones to carry tight $F3_s$ minor folds, which are more nearly reclined than those in the core zones. This is not always so - e.g. contrast Plates 8.1(b) and 8.1(g), both showing no particular sense of vergence, with Plate 8.1(e). In detail, the axial planar fabric is a crenulation of both the coarse migmatitic lits and the finer schistose matrix (Plate 8.3(a)); it is almost penetrative on the NW and SE limbs of the major fold (Plate 8.3(b), 56276), but open crenulations are seen in the SW core zone (Plate 8.1(e)). $F3_s$ folds in the psammites tend to have rounded hinges (Plate 8.1(c), (f)) and to carry an axial planar shape fabric defined by quartz aggregates a few millimetres across. This is an LS fabric with the L component plunging consistently to about 160° (Fig. 8.3); in the reclined SW core zone, fold axes are parallel to this lineation. In the central, gently plunging zone, fold styles in both core and limbs are similar to those in the reclined zone, in terms of tightness, profile, and nature and orientation of the axial planar fabric. The folds, however, have variable plunges, usually pitching <45° on $S3_s$ (e.g. Plate 8.1(c)), and the $L3_s$ extension lineation is oblique to the fold axes. The fold axes steepen in zones, i.e. moving either NW or SE away from the centre, alternating gently and steeply-plunging zones are seen, with the latter progressively increasing in width at the expense of the former.

In the area south of [975283], tight minor folds of uncertain age are common on axial planes dipping about 40/140 with steeply south-plunging axes. These are similar in style to $F3_s$, some have an axial planar shape fabric, and they show Z-profiles. Further south, M and W folds are common, and at least one major closure (south-plunging synform with wavelength >50m, amplitude >200m) is seen in massive psammitic beds at [978282]. In style and orientation the associated minor folds look like $F3_s$; however, the characteristic $S3_s$ shape fabric is not seen. This may be a complementary synform to the Sguman Coinntich antiform.

Interference patterns involving $F3_s$ are rare but widespread. Plate 8.1(b) shows a reclined $F3_s$ fold, with two earlier, nearly isoclinal folds of bedding passing round it, all being almost coaxial. The cleavage in the core of the $F2_s$ fold is also a crenulation, of micas and migmatitic lits; the $F1_s$ fold has a penetrative foliation with small-scale migmatitic segregation. Minor S-profile folds of thin lits probably belong to the adjacent $F2_s$ fold rather than the $F1_s$. In Plate 8.1(f), isoclinal $F2_s$ folds in Morar Division psammites are refolded by reclined $F3_s$. Plate 8.1(d) shows (gently plunging) $F3_s$ folds of the platy slide fabric, and in thin section, this ($S2_s$) fabric is itself a crenulation cleavage (Plate 8.3(c)). These suffice to demonstrate that the Sguman Coinntich fold is at least a third structure.

At [978293], on the long limb of an intermediate-scale $F3_s$ fold, textures reminiscent of Mallaig S4 (section 3.3.2) occur. A coarse mica fabric, which is probably $S3_s$ or a composite $S2_s$-$S3_s$, is overprinted by narrow anastomosing zones of grain size reduction. The fine grain size seems at odds with the general recrystallisation and pegmatite emplacement of $MS3_s$ age (sections 8.3, 8.4). They could be regarded as late $D3_s$ in age or may be related to some later structure (e.g. the synform SE of the pelite, if it is actually post-$F3_s$; cf. the zones of late $D3_a$ strain at Acharacle, section 6.2.1).

8.2.2 $F2_s$

This deformation gave rise to the main pre-$F3_s$ fabric of the area, and involved the formation of a major zone (or zones) of high strain, probably equivalent to the Sgurr Beag Slide.

Structural relations are best observed in the SW core of the Sguman Coinntich fold. There, a traverse from NE to SW, in relatively open, reclined $F3_s$ fold cores, reveals a strain gradient related to the main pelite/psammite boundary. Within the main pelite outcrop, the fabric which is folded is a strong crenulation cleavage of an earlier coarse mica fabric and migmatitic lits (e.g. Plate 8.3(c), 56233). Rare tight-to-isoclinal folds occur, deforming bedding and migmatitic banding (Plate 8.1(b), (g)). Towards the slide, $S2_s$ becomes penetrative and lits are strongly augened. The slide itself presents a sharp boundary (see below), and is followed by very finely banded psammites with a strong, platy shape fabric. This fabric becomes less intense to the SW, and about 50m from the boundary, isoclinal $F2_s$ folds appear (Plate 8.1(f)). The psammites there are flaggy rather than platy (banded on a centimetre or so) with tight intrafolial folds (e.g. Plate 8.2(b)), 150m from the slide). About 300m from the boundary, $F2_s$ deformation becomes more heterogeneous, and flaggy zones alternate with zones of relatively open folding (Plate 8.2(c)). This pattern is typical of that at the Sgurr Beag Slide (e.g. Rathbone 1980, Rathbone & Harris 1979; sections 5.2.1, 6.2.2, 7.4). The alternating flaggy/open-folded pattern is particularly reminiscent of Kinlochourn, since although the open $F2_s$ folds here are reclined, while the open $F2_k$ folds there are gently plunging, the $F2_s$ axes could have been rotated during the $D3_s$ deformation, which is more penetrative here than is $F3_k$ west of the slide at Kinlochourn.

The slide clearly predates the Sguman Coinntich fold; this is confirmed by following the contact, where the above-mentioned features are seen all round the pelite. The slide is exposed on the NW limb at [971298] (Plate 8.2(a), (d)). It consists of a zone a few metres wide of coarsely recrystallised micaceous rock, which is clearly highly deformed migmatitic pelite, with feldspar porphyroclasts and quartz ribbons in a muscovite-rich matrix. This contains lenses a few metres long of both migmatitic pelite (white-weathering) and psammite (dark-weathering). Several thin bands of similar platy rock cut into the overlying pelite (marked in Plate 8.2(a) by lines of heather). A similar pattern is seen at the SE boundary, though less well-exposed. The immediately adjacent psammites are very platy, with a strong shape fabric (Plate 8.4(a), (b)), which is folded by minor $F3_s$ structures.

Further south-east in the psammites, occasional early isoclines are involved in $F3_s$ interference patterns; these and the crenulated fabric may be of $F2_s$ age. A number of Lewisianoid inliers lie about 300m east of the pelite boundary at [986294], and might be expected to mark the position of another slide. However, no significant increase in platiness is seen towards them, and no metamorphic contrast is seen (this may partly be due to overprinting of shape fabrics by $S3_s$, since clearly a shape fabric in psammites is more easily modified by deformation or metamorphism than the crystallographic and location fabric in a pelite). In any case, the major slide seems to occur at the psammite/pelite boundary.

8.2.3 $F1_s$

All earlier structures will be included in this category. No major structures were seen; minor structures include the pre-$F2_s$ fabric in (56233), seen in Plate 8.3(c), and the migmatitic banding, which presumably had some fabrics associated with it. An $F1_s$ fold is seen in Plate 8.1(b); it is isoclinal, and has an axial planar fabric which is apparently penetrative. Quartzofeldspathic segregations are developed in this fabric, possibly indicating that migmatisation was $MS1_s$ or $MP1_s$. In the psammites, no $F1_s$ structures were recorded; they probably would not be recognised in the study area, with its high $D2_s$ and $D3_s$ deformation. In the least strained psammites at [959290] (Plate 8.2(c)), a weak mica fabric is folded by $F2_s$, but no shape fabrics occur. This may represent $S1_s$, but it could be argued that it is of sedimentary origin.

8.2.4 Later structures

Open, upright folds on axial planes trending 150-160° are common; these deform the $S3_s$ fabric but are coaxial with the $L3_s$ lineation (and with $F3_s$ folds where these are steeply pitching). Folds of this age are probably responsible for the swings in strike of the pelite boundary and of $S3_s$ (Fig. 8.2). Fig. 8.3 shows the approximate orientation of the main open antiform south of the Sguman Coinntich summit - the axial plane dips at about 75/228, with axis plunging 30/145; the maximum interlimb angle is about 50°. Crenulations are open and sharp-crested with strained micas.

The variation in dip of $S3_s$ and plunge of $L3_s$ cannot be explained by these folds - a structure with a gentle, NE-SW axis is required. Some minor folds of this orientation do occur, e.g. in Plate 8.1(c) (note that the upper open fold deforms the $S3_s$ fabric, which is axial planar to the lower fold), but no mappable structures are present.

8.3 Metamorphism

This will be divided into pre-slide, syn-slide and post-slide events. Precise textural evidence is rather scanty, and reference will be made to section 8.4 on migmatisation to help define conditions in the psammites.

8.3.1 Pre-slide metamorphism

It is clear that in the pelite, the coarse, migmatitic texture pre-dates the $D2_s$ deformation; lits are crenulated and augened (Plates 8.1(b), 8.2(d), 8.4(c)), and coarse quartz and feldspar is strained (Plate 8.3(c)). They are similarly deformed by $F3_s$ structures (Plates 8.1(a), 8.3(a),(b)). Garnet is also augened by $S2_s$, although in the absence of inclusion trails, it is not always possible to state conclusively that a given garnet is entirely pre-$F2_s$ (cf. Acharacle, comments at the end of section 6.4.2). Garnets in pelites (and some amphibolites) have a marked concentric, quartz-rich inclusion zone (Plate 8.3(d)). A typical pelite assemblage would be quartz-muscovite-garnet-biotite-plagioclase (An_{20-30})-ilmenite. Garnetiferous amphibolites (e.g. 56277) consist largely of brown hornblende and garnet, with minor quartz, plagioclase (An_{50}), sphene and ilmenite; Fe-rich epidote occurs as inclusions in the hornblende. The pelite assemblage would be stable from the garnet to the sillimanite-orthoclase isograds. However, the coarse, migmatitic textures suggest fairly high grade, i.e. kyanite or higher. The brown hornblende, presence of large garnets and absence of biotite, and the calcic plagioclase in the amphibolite also suggest kyanite grade or higher. P,T estimates based on mineral chemistry (section 15.2.3(ii)) of 600°C at 6.5kb put the rocks well into the kyanite zone.

The Morar Division rocks are generally lacking in index minerals. Their general coarseness is comparable to that of other Morar Division psammites; plagioclase is high oligoclase, biotite is abundant while chlorite and zoisite are absent, probably indicating at least lower amphibolite facies. At one locality ([973280], 56178), tiny garnets are included in plagioclase; these may be $MS1_s$, but could be later. A characteristic feature of the psammites is the presence of Fe-rich clinozoisite or Fe-poor epidote, not apparently derived from retrogression of plagioclase. These are not found in the Glenfinnan Division, except as a retrogressive phase; this presumably indicates lower grade for the Morar Division rocks (i.e. implying that a psammite, which in the Morar Division has An_{25} plagioclase and epidote, would have a more calcic plagioclase in the Glenfinnan Division).

8.3.2 Syn- to post-slide metamorphism

Precise $MS2_s$ metamorphic grade is difficult to estimate. Micas in the $S2_s$ cleavage and crenulations are unstrained, and new biotite has grown, indicating at least biotite grade. Feldspar at An_{20-30} is stable suggesting higher (garnet or staurolite) grade. No new garnet was recognised, but textural relations would permit some $MS2_s$ growth at the rims of earlier garnets (cf. Plate 8.3(b), (c), Fig. 6.5(c)). Some $MS2_s$ or $MP2_s$ migmatisation occurred (section 8.4.2), but not to the same extent as during $MS3_s$.

The same textural comments apply to $MS3_s$-$MP3_s$. Garnets are augened, but micas and matrix quartz show no strain features. Migmatisation was more widespread, which could indicate higher temperatures, but other explantations such as a difference in the mechanism of deformation, or even that $D2_s$ was shorter-lived (in time) than $D3_s$, are possible. Where this "late migmatisation" has occurred, a notable feature is the abundance of epidote in biotite-rich selvedges. Since epidote does occur in non-migmatised psammites (e.g. 56282), it is not clear if this has formed due to the migmatite-forming reactions (presumably giving less calcic feldspar in the lits) or has merely been concentrated with biotite as a restite (see section 8.4.2). In terms of degree of "late migmatisation" and $MP3_s$ coarsening, the characteristics of this area lie between those of the slide zone at Kinlochourn and at Acharacle. The effects of migmatisation seem to be stronger in the south-east, and, this being so, it may be inferred that a "front of migmatisation" runs close to the south-east boundary of the area mapped - suggesting that $MS3_s$ grade may have been in the upper garnet to lower staurolite zones (see section 6.4.2 for an expansion of the reasoning involved in reaching this conclusion).

The later, coaxial refolding clearly took place at lower grade - micas are bent and strained with no recovery in even the tightest crenulations.

8.4 Migmatisation

Migmatisation will be considered under two headings - early migmatisation, restricted to the Glenfinnan Division, and later migmatisation, developed throughout the area.

8.4.1 Early migmatisation

The Glenfinnan Division here consists of coarse, homogeneous, lit-par-lit migmatitic pelitic gneiss. The lits (Plates 8.1(a), 8.2(d), 8.3(a)) were augened or folded during $D2_s$ and $D3_s$, and their constituent quartz and feldspar is deformed (Plate 8.3(b), (c)). They consist of coarse (>1cm) quartz and oligoclase (about An_{25}), with the latter forming 50% of the vein. Where they are undeformed, individual veins or lits range in thickness from 1cm to (rarely) 10cm, and extend for 20-40 times their thickness before lensing out (e.g. Plate 8.1(e), (g)). The degree of segregation varies - e.g. the semipelitic bands in Plate 8.1(b) are almost non-migmatised, while the pelite in Plate 8.1(e) consists entirely of lits and pure mica/garnet layers. The analysed sample (56180) came from the latter locality (section 16.2). In Plate 8.1(a), one band is highly migmatised, while the adjacent one is unaffected. The matrix in highly migmatised rocks consists only of the garnet/mica selvedge - in effect, there is no palaeosome, only neosome consisting of leucosome and melanosome (sections 2.2.2, 17.1). However, the typical pelites have a matrix which is a finer-grained version of the migmatitic texture - i.e. thin lenses and stringers of quartz and oligoclase lie concordantly in a schistose matrix (Plate 8.3(a)). These finer lenticles lie in the axial planar foliation to the $F1_s$ fold in Plate 8.1(b), which may indicate an $MS1_s$-$MP1_s$ age for the migmatisation. Certainly, there is no evidence for post-migmatisation, pre-sliding deformation of the lits; the quartz inclusion-rich zone in the (chemically unzoned) garnets may correspond in time to the migmatisation (Plate 8.3(d) - cf. sections 5.3.2, 5.4.1). Their likely mode of origin is discussed fully in Chapter 17, but lit composition and the estimated temperatures make a subsolidus origin most likely.

8.4.2 Later migmatisation

A broad definition of migmatisation is used here, so that some phenomena which relate more to pegmatite injection, but which are part of a general pattern of metamorphism, segregation and injection, are not excluded. The reasoning and justification behind this is developed fully in section 6.4.2. Late migmatisation is largely restricted to the psammites - this is a general pattern, even within the Glenfinnan Division; remobilisation of a pelitic migmatite takes place only when the texture has previously been broken down by deformation. In this area, conspicuous, coarse, K-feldspar-bearing pegmatite dykes are restricted to the SE, and are late and cross-cutting ($MP3_s$).

Veins which are probably locally derived show a close association with fold limbs or axial planes, have compositions similar to that of their host, and are bordered by a biotite-rich selvedge if the psammite is micaceous (Plate 8.4(d)). They often have irregular shapes and lens out laterally. Composition tends to reflect that of the host rock, in so far as microcline is only present when the psammite also contains it. Plagioclase is usually similar to that in the host rock (56280, 56281). Schlieren of biotite with abundant epidote inclusions are common. This pattern does not rule out any of the possible modes of origin (i.e. injection, partial melting or sub-solidus metamorphic differentiation), although the dependence of host-rock composition means that injection could only be considered in conjunction with one of the others, giving a hybrid origin. The similar plagioclase composition in host and lit could result from melting at high water pressure, or from sub-solidus processes (cf. Yardley 1978; section 17.1). The veins are similar in appearance and composition, regardless of age, but are much more abundant during $MS3_s$-$MP3_s$; they also become more abundant to the south-east, concomitant with a coarsening of psammite textures.

Plate 8.1(f) shows an $MS2_s$ pegmatite, emplaced parallel to an $F2_s$ axial plane, cutting folded bedding, but itself carrying an $S2_s$ fabric. In Plate 8.1(d), a lit, which is still quite coarse and undeformed, lies in the $S2_s$ platy fabric and is folded by $F3_s$; in extreme cases, coarse, late veins produce a lit-par-lit psammitic migmatite. Plate 8.4(d) shows veins developed axial planar to $F3_s$ folds in psammites SE of the pelite at [974290]. In the degree of late migmatisation, the characteristics of this area lie between those of Kinlochourn and Acharacle.

8.5 Discussion

Except for the fact that the migmatitic pelite has no outcrop continuity with the main Glenfinnan Division, the situation at Sguman Coinntich is similar to that at Kinlochourn and Acharacle. The Morar Division was metamorphosed at garnet grade or higher; the Glenfinnan Division was migmatised. They were then juxtaposed by movement on a major slide, and tight folding with local migmatisation followed. It was shown in section 8.2.1 that the Sguman Coinntich structure must be an antiform, so the Glenfinnan Division cannot simply have been brought down in a klippe (cf. Clifford 1958a). Clifford's NW-SE section has a number of other inconsistencies - e.g. the Beinn Bhreac fold east of Coire nan Gall would appear to open upwards initially, before closing off as an antiform; Fleuty (1974) has remapped the Beinn Bhreac area, and shown that the fold is a synform.

The migmatitic pelites on Sguman Coinntich therefore underlie the Morar Division, in contrast to the situation from Kinlochourn southwards. This might be taken to indicate that they are not part of the same nappe as the main Glenfinnan Division outcrop. However, the typical slide fabric is still present (suggesting that the migmatites do not represent deep-seated Morar Division rocks), and the "shells" of psammite, surrounding the pelite, are identical to those surrounding the Beinn Dronaig pelite (A.L. Harris, verb. comm.). The latter apparently rests on a synform (Sutton & Watson 1955), probably on the Sgurr Beag Slide; this suggests that the psammite surrounding the Sguman Coinntich pelite is at a high structural level in the Morar Division. Assuming that the Glenfinnan Division was thrust over the Morar Division, a sense of "facing" of post-slide structures can be determined by treating the slide as if it were a stratigraphic boundary (cf. section 2.2.1). On this basis, the Sguman Coinntich antiform is downward-"facing" across the slide, while the Beinn Dronaig synform is upward-"facing". It is noteworthy that in Glen Shiel, a few hundred metres west of the Sgurr Beag Slide, structures similar to $S3_s$ were observed which are genuinely downward-facing in right-way-up Morar Division psammites. These points could be reconciled by postulating an additional, recumbent isoclinal fold, post-dating the slide, with the Sguman Coinntich/Glen Shiel and Beinn Dronaig areas being on opposite limbs, at the present level of exposure. Clearly, detailed mapping of the ground between Sguman Coinntich and Beinn Dronaig would be required to clarify this situation.

Regarding the origin of the Sguman Coinntich fold itself, Clifford (1958a) suggested that its variation in plunge was due to the later, NW-SE folds. However, they clearly are not tight enough to effect a near-180° rotation of the $F3_s$ axes. From Fig. 8.3, it is clear that the variation in $F3_s$ axes is within unchanging $F3_s$ axial planes - if the structure was due to interference, then $F3_s$ would have to be the second, and not the first fold phase. Even then, one would expect to find remnants of a vertical, NW-SE cleavage in the central region of the fold, at a high angle to bedding, and overprinted by shallower $S3_s$. The most acceptable explanation is that the curvilinearity is an inherent feature of the $F3_s$ deformation - e.g. due to inhomogeneous flattening, or a superimposed plane strain amplifying initially weakly-curved axes (Sanderson 1973). The strong LS fabric might favour the latter mechanism, with nearness to the Moine thrust belt suggesting the operation of simple shear.

CHAPTER 9

Ben Klibreck

9.1 Introduction

Ben Klibreck lies in central Sutherland, some 30km east of the Moine Thrust. Rocks similar to those of the Morar Division dip steadily southeastwards from the Moine Thrust, although in their eastern part, Lewisian inliers concordant with foliation become abundant. The main western scarp of Ben Klibreck is formed by rocks of the Loch Coire migmatite complex, part of the larger east Sutherland migmatite complex (Fig. 9.1). These are similar, at least on Ben Klibreck, to the pelitic gneisses of the Glenfinnan Division further south (Read 1931, Johnstone 1975), and overlie the Morar Division rocks, probably on a structure analogous to the Sgurr Beag Slide (section 9.4.1). Both divisions show widespread granite and pegmatite injection. The aim of this study was to determine what connection, if any, this injection had with the regional migmatisation of the Glenfinnan Division, and to compare this area with the superficially different rocks further south (cf. discussion by Brown *in* Butler 1965).

The area was first described by Read (1931), who cited Ben Klibreck as a type section for the Loch Coire migmatite complex. The migmatites were regarded as having formed by injection of granitic material into a dominantly Moine host, consisting of psammite in the west, and pelites and semipelites in the east. A number of outcrops of "rocks of Lewisian type" were recognised; these were compared petrographically with the Lewisian inliers of Ross-shire, although it was noted that, in this area, no structural or metamorphic distinction could be made between Lewisian and Moines. Read suggested that both Moine and Lewisian had been affected by a pre-Torridonian metamorphism.

Modern work has centred on the relationship of deformation and metamorphism in the Morar Division to that in the Moine Thrust

Belt, and the origin of the Loch Coire migmatites. Soper (1971) and Soper & Wilkinson (1975) described a four-fold sequence, in which D1 was restricted to the mylonite belt, D2 comprised a ubiquitous ESE-dipping LS fabric with tight, reclined F2 folds, D3 was represented by more open, westerly-overturned folds, and D4 formed box folds and kink bands. All these were interpreted as post-Arenig, i.e. Caledonian. A suite of pegmatites and minor intrusions (the Vagastie suite), which was rodded by D2, was tentatively correlated with the Carn Chuinneag granite (Shepherd 1973). Soper & Brown (1971) described a west-to-east increase in metamorphic grade, from biotite in the mylonites to sillimanite in the Loch Coire complex (based on pelites and calc-silicates). The main recrystallisation was MP2, and was correlated east into the migmatites, and south as far as Carn Chuinneag. Kyanite and sillimanite were oriented parallel to the L2 lineation, but the possibility of passive rotation was not, apparently, considered. The western and lower limit of the migmatite complex was taken to be the Lewisian inlier at the foot of the main scarp (cf. Figs. 9.2, 9.3). To the west, a "zone of veins", of aplite, pegmatite and quartz, was defined; F2 folds tightened towards the migmatites, and the lack of rodding of <u>lits</u> in the higher part of the complex was taken to indicate an MP2 origin, although the possibility of some late-D2 homogeneous flattening was allowed. The apparent metamorphic inversion was considered to be primary, possibly due to fluid movement up a pelite horizon in the migmatite complex.

Brown (1967, 1971) concluded from geochemical evidence that the migmatites were metasomatic. The key features were the more sodic nature of the migmatites in relation to the unmigmatised Morar Division, and the wide spread of vein compositions away from the ternary minimum in the Q-Ab-Or system. This contrasted with the work of Butler (1965), who found sodic metasediments to be common in the Morar Division of Ardnamurchan; Brown suggested that their absence in the north was due to greater sedimentary maturity, bearing in mind the general northward palaeocurrent directions found in the Morar Division (e.g. Richey & Kennedy 1939).

In the light of recent work on the interpretation of deformation textures (e.g. White 1977), and the regional structure of the Moines (e.g. Tanner et al. 1970, Rathbone & Harris 1979, Rathbone 1980), many of the criteria used in reaching the above conclusions must be regarded as suspect. In particular, if the Loch Coire complex consisted in part of previously gneissose Glenfinnan Division, and the MP2 metamorphism has merely recrystallised earlier stable mineral assemblages, then a model in which the metamorphic inversion is due to early thrusting of high grade rocks over lower, would be equally consistent with the evidence given by the above authors. The presence of Lewisian inliers, platy zones and Glenfinnan Division lithologies overlying Morar Division suggests the presence of slides, possibly analogous to the Sgurr Beag Slide, in this region. Two sections were chosen for investigation: a traverse up the main western slope of Ben Klibreck (Fig. 9.2), and one across the isolated body of migmatitic pelite to the southwest (Traverse 2, Fig. 9.1).

In the first section, the rocks dip steadily southeastwards, and consist mainly of Morar Division psammites, with subordinate garnetiferous semipelites. These psammites are coarse and quartzose, although locally arkosic, and where deformation is low form beds 10cm to one metre thick. A number of concordant Lewisian inliers occur. The two thicker ones (at [564305] and [572300]) contain a variety of lithologies, dominated by striped plagioclase-hornblende-biotite-quartz gneiss of dioritic to tonalitic composition (e.g. 56614), although more mafic hornblende schists and some coarse acid gneisses are also present. The two smaller inliers (at [567305] and [577300]) are poorly exposed, but appear to contain a greater proportion of mafic material. A coarse, granular metabasic body or "epidiorite" occurs near Loch an Tairbh; the contacts are not exposed, but its topographic expression suggests a boss-like form. In the southeast corner of Fig. 9.2, migmatitic pelite abruptly succeeds Morar Division psammites; this is a typical coarse, garnet- and biotite-rich Glenfinnan Division pelite, containing concordant quartz-oligoclase lits. It becomes semipelitic upwards, and on the higher slopes, micaceous psammites and semipelites dominate. A Lewisian inlier occurs in these rocks; it consists mainly of feldspathic hornblende gneiss and quartzofeldspathic gneiss. Granite and pegmatite veins are common throughout the eastern half of the map; they are usually K-feldspar-bearing.

Regionally, it can be seen that these Glenfinnan Division rocks form part of the Loch Coire migmatite complex, and that the Lewisian inlier at [572300] is the southern continuation of the Loch Naver inlier (Fig. 9.1). A lenticular body of "veined pelite" is shown on the I.G.S. Sheet 108 at [560270]. Traverse 2 passed from Morar Division psammite, similar to that described above, into migmatitic pelites, petrographically identical to those on Ben Klibreck. Exposure is poor, and neither the boundaries of the pelite, nor the presence of a Lewisian inlier, could be verified in the time available.

The structure and metamorphism will be described in the two succeeding sections, and then particular topics (slides, Lewisian, regional migmatites, granites and pegmatites) will be enlarged upon.

9.2 Structure

The dominant structures are of local F2 age ($F2_b$); these will be described first, then evidence adduced for an earlier deformation. Late structures are negligible, being restricted to a few open folds in the migmatitic pelites.

9.2.1 $D2_b$

$F2_b$ structures occur throughout the area. In the west, they are represented by alternating zones of tight-to-open reclined folds, with wavelength up to a few metres (Plate 9.1(a)), and flaggy to locally platy rocks, with rare tight S- or Z-folds (Plate 9.1(b)). These zones correspond to the cores and limbs respectively, of larger $F2_b$ folds with wavelength about 100m. They control the topography, with folded zones forming small scarps, while flaggy zones form poorly exposed dip slopes. Axial planar

$S2_b$, in the folded zones, and foliation (composite bedding-$S1_b$-$S2_b$), in the flaggy zones, dip uniformly to the southeast (Fig. 9.4). Fold axes generally pitch steeply SW on $S2_b$, sub-parallel to mineral and intersection lineations (Plate 9.2(a)). A strong LS quartz grain aggregate shape fabric is parallel to $S2_b$, with L parallel to the fold axes in most cases. Towards the southeast, folds become tighter (Plate 9.1(c)), and the broad, open-folded zones of the west are absent. In the main scarp, above the Loch Naver Lewisian, $F2_b$ folds are rare and sub-isoclinal. In the Glenfinnan Division, $F2_b$ folds reappear, although they are never as open as those of the western area (Plate 9.1(d)). Throughout the whole area, $F2_b$ orientation is remarkably constant (Fig. 9.4 - $F2_b$ axial planar structures have been distinguished from possibly-composite foliations in the flaggy zones). $S2_b$ dips average 25/120, while those in flaggy zones average 29/117. The trend of $F2_b$ fold axes is essentially parallel to that of other lineations (mean 18/156, as against 22/154), while the shallower plunge reflects the shallower dip of $S2_b$. A rough test of the significance of the steeper flaggy foliation was made by applying the Mann-Whitney U test to the collected dips (cf. section 16.2.3). The conclusion was that there is no significant difference in dip at the 95% confidence level. This also applies to the fold and lineation plunges.

The $S2_b$ foliation is generally penetrative in the Morar Division, but locally is a crenulation cleavage (Plate 9.2(b)). In the Glenfinnan Division, the gneissic foliation is deformed by $F2_b$ (Plate 9.1(d)), and on limbs, the foliation is probably composite in origin. Many granitic or pegmatitic bodies carry an LS fabric parallel to $S2_b$. However, it is clear from their detailed relations that most of them postdate the $F2_b$ folds (section 9.4.4). Their foliation probably originates from a period of relatively homogeneous straining, late in $D2_b$ (at least, before final $MP2_b$ annealing). The earliest granite sills carry a strong LS fabric, with quartz aggregates showing axial ratios of the order 2:1:0.5, with L parallel to the local $F2_b$ axes. It is possible that these were hotter than the surrounding Moines, and so acted as ductile horizons on which large displacements took place. However, features such as a fissile margin (cf. the microdiorite suite, Smith 1979) are absent, so the sills may have deformed relatively homogeneously with their host rocks. If so, much of the $D2_b$ strain, and $S2_b$ fabric development, postdates $F2_b$ fold formation. This would account for rotation of $F2_b$ fold axes into parallelism with the extension direction, even in areas where the folds are still relatively open. Such a deformation history would accentuate the contrast in bed thickness between fold cores and limbs, as observed in the west.

Quartz grains in the granite sills show rectangular outlines (and some ribbons) parallel to the foliation, which are remarkably similar to those depicted by White (1977) for grain boundary sliding in a simple shear environment. However, these grains are much coarser (Plate 9.2(c)), which should favour intracrystalline deformation (some has occurred, as indicated by undulose extinction bands in the quartz grains). Psammitic country rocks show coarse, equant recovery textures. Some of the later, cross-cutting, K-feldspar-rich veins carry an LS fabric with L significantly anticlockwise of $L2_b$ (e.g. 25/125 vs. 16/155 at [565306]). This may indicate a change to a more westerly direction of translation late in $D2_b$.

Large scale $F2_b$ structures cannot be defined with certainty. Soper (1971) suggested that the Loch Naver inlier is a D1/D2 interference structure. Certainly the westernmost inlier on Fig. 9.2 lies in a fold core - M and W folds are abundant within it, Lewisian lithologies are symmetrical, and $F2_b$ folds change vergence across it. At its northern end it is a synform; however, fold axes probably curve through the horizontal to give the lenticular outcrop on Fig. 9.1, so the overall structure could close either upwards or downwards. The southern tip of the Loch Naver inlier is lithologically very similar, and a zone with abundant M and W folds lies midway between, so correlation of the two is reasonable. The two smaller inliers are both rich in mafic rocks, so they, too, could be correlated. Also, all the inliers have a strong $S1_b$ fabric at their margins; this fabric can be recognised in the absence of Lewisian, and may represent a horizon at which Lewisian has been slid out (section 9.2.2). One such zone occurs in the Allt a' Mhuilinn, and may correspond to the second Lewisian inlier. This horizon may also be represented by the platy zone west of Loch na Glas-Choille. The few $F2_b$ folds in the main scarp have a consistent westward vergence; however, sufficient evidence could not be found to determine if the Loch Naver Lewisian occupies a fold core. Two possible sections are shown in Fig. 9.3. Section (a) assumes that the Loch Naver Lewisian occupies a fold core; the two small inliers and the platy zones then join up to form a higher horizon than the Loch Naver Lewisian, and the westernmost inlier closes downwards; it could equally well be drawn to close upwards, by changing all antiforms into synforms, but the regional structure is more likely to be overturned to the west. The first section is simpler, in that it reduces the number of levels at which Lewisian inliers lie; it also fits better with the regional map (Fig. 9.1). The migmatitic pelites of Cnoc Sgriodain lie along strike from a synform, and could be an outlier resting on the main slide. Traverse 2 showed a series of tight, Z-profile $F2_b$ folds in the psammites, plunging at low angles to the south or SSW on the main ridge; 200m to the north, fold plunges steepen to values such as 20/155, on axial planes dipping about 25/125. This would be consistent with a tight, downwards-closing, keel-like structure.

It may be significant that the Crask Lewisian to the west of Cnoc Sgriodain lies along strike from a platy zone; on section (a), it could lie at the same horizon as the Loch Naver Lewisian. On either section, it could also correspond to the folded zone near the A836, and hence lie in a fold core, probably at a lower structural level. In summary, regional $F2_b$ structures are probably generally reclined, but doubly-plunging, and overturned to the west on a large scale. The available evidence favours section (a), but much wider mapping would be required to define the structure with any certainty, and even then, the poor exposure might prevent a unique solution being obtained.

9.2.2 $D1_b$

All pre-$F2_b$ structures will be included in this section. It is probable, particularly in the Glenfinnan Division, that more than one phase of deformation is presented.

In the Morar Division, $S2_b$ is locally a crenulation cleavage (Plate 9.2(b)), and refolded isoclines occur in the $F2_b$ fold zones (Plate 9.1(e)); these are coaxial with $F2_b$.

At the margins of Lewisian inliers, a strong platy fabric is seen, which is folded by $F2_b$ structures (Plate 9.1(f)); this extends for perhaps 50m into the psammites, but dies out rapidly into the Lewisian. A strong LS fabric, with quartz ribbons several centimetres long, and feldspars thoroughly recrystallised and elongated (Plate 9.2(d)), is present in acid gneisses near the boundary. Within the inlier, tight $F2_b$ folds of 0.5-1.0cm banding occur (Plate 9.3(b)); a weak quartz-feldspar shape fabric and a hornblende orientation fabric go round these folds (e.g. 56603); banding is locally transposed into $S2_b$. The feldspathic hornblende gneiss is coarser grained (0.6mm vs. 0.3mm), and more coarsely banded (1cm vs. 2mm) near the centre of the inlier. Elsewhere in the Morar Division, zones of very platy psammites occur (Plate 9.1(g)). The fabric is much more intense than any associated with $F2_b$ structures, and may be related to sliding during emplacement of the Lewisian - the cross sections favoured above would put the platy zones at the same horizons as Lewisian inliers.

In the Glenfinnan Division of Ben Klibreck and Cnoc Sgriodain, migmatitic banding and as associated foliation are folded by $F2_b$ structures, and in both areas, interference patterns occur in which $F1_b$ also folds migmatitic banding. A strong platy fabric is present throughout the psammites of the main scarp, although it is obscured by intense late granite and pegmatite intrusion.

This may be related to some analogue of the Sgurr Beag Slide (cf. Rathbone 1980, Rathbone & Harris 1979), which has emplaced the

Glenfinnan Division onto the Morar Division, and which is probably of $D1_b$ age (section 9.4.1).

Psammites and acid gneisses in the platy zones show rectangular quartz grains, similar to those developed in the granite sills during $D2_b$. This could indicate grain boundary sliding under simple shear, although again the grain size is rather high (0.2mm - Plate 9.2(d), (e)). Elsewhere in the Lewisian and Morar Division, the rocks have a coarse polygonal texture. Possibly a combination of high strain rate with channelling of fluids along the slide zones would permit grain boundary sliding to dominate locally, even at these grain sizes.

9.3 Metamorphism

Lithologies suitable for defining metamorphic grade are scarce in this area, and late recrystallisation has made textures difficult to interpret. The tacit assumption of previous workers such as Read (1931) and Soper & Brown (1971) that the peak of metamorphism, as indicated by index minerals, was synchronous with the last widespread recrystallisation of the matrix, is not considered to be valid in an area where there is a strong possibility of polymetamorphic or polyorogenic events. Since all the major units are tectonically bounded, each will be described separately, before a final synthesis is made.

9.3.1 Morar Division

Rare garnets, present in both psammites and semipelites, and relatively anorthitic plagioclase (An_{20-30}) indicate at least garnet grade for $MP1_b$ (Plate 9.2(f)). The main matrix coarsening was $MP2_b$ in age, as micas in $F2_b$ crenulations are recrystallised, and quartz and feldspars show unstrained, polygonal textures (Plate 9.2(b)). There is an eastwards and possibly upwards increase in grain size within the Morar Division. West of Loch na Glas-Choille, the quartz matrix has a grain size of about 0.2mm; east of the first Lewisian inlier, it increases to about 1mm (Plate 9.2(g)). In the main scarp, quartz grains are very coarse (2mm) and equant, but microcline is elongated 2 or 3:1 in the foliation (58143). Since no strain features are evident, in rocks with high $D2_b$ deformation, this recrystallisation must be $MP2_b$. It may be due to the extensive syn- to post-$D2_b$ intrusion of granites. The age of all garnets could not be determined with certainty, due to their small grain size in relation to the matrix, but some certainly pre-date $S2_b$ (Plate 9.2(f)). The epidiorite body carries a $D2_b$ fabric, so its assemblage of green hornblende and An_{35} plagioclase is probably of $MS2_b$-$MP2_b$ age; it would be stable from garnet to about kyanite grade. It shows some retrograde features, with plagioclase zoned from An_{50} to An_{30}, and ilmenite mantled by sphene, but these ostensibly high grade relics could be igneous in origin. Biotite inclusions in the large quartz grains of (56621) - Plate 9.2(g) - indicate that this is a genuine $MP2_b$ coarsening, and not merely inheritance of an earlier coarse-grained texture.

9.3.2 Glenfinnan Division

The Glenfinnan Division is immediately distinguished by its coarse, migmatitic texture. The migmatitic lits certainly pre-date $D2_b$, and probably $D1_b$ as well (section 9.4.3); they are identical to those found elsewhere in the Glenfinnan Division, and must indicate middle to upper amphibolite facies. The dominant foliation in the pelites is probably a composite $S1_b$-$S2_b$; extreme post-tectonic coarsening often makes garnet textural relations ambiguous, but some are certainly augened (Plate 9.2(h)). They are probably approximately coeval with migmatisation, and so may be pre-$D1_b$. Glenfinnan Division psammites have suffered $MP2_b$ recrystallisation, comparable to that of the higher Morar Division psammites.

9.3.3 Lewisian

The main information about metamorphism in the Lewisian comes from the hornblendic rocks. True hornblende is present in feldspathic gneisses; it shows dark green to brown-green pleochroism, and is associated with An_{30} plagioclase. In more basic rocks, plagioclase ranges up to An_{50}, while in quartz-bearing rocks, Fe-rich epidote is common. Hornblende is often randomly oriented, and the matrix unstrained, so an $MP2_b$ origin is likely. A new differentiated $S2_b$ fabric, with stripes rich in unoriented hornblende crystals, appears to be formed by segregation processes rather than transposition, e.g. in (56615).

9.3.4 Discussion

All the rocks are clearly affected by an amphibolite-facies $MS2_b$-$MP2_b$ metamorphism, with an eastwards increase in grain size of the resultant textures. This may be due to local heating by the intense swarm of granites and pegmatites. No definite $MP2_b$ garnet was observed, although many possible examples occur. However, the epidote-free basic assemblages, with their calcic plagioclase, suggest staurolite or kyanite grade. The Lewisian inliers were emplaced, possibly on slides, during $D1_b$. No direct evidence for $MS1_b$ grade is available, but the strong ductile fabrics in acid gneisses at the margins of inliers indicate that the slides were syn-metamorphic shear zones, rather than cataclastic fault zones. Prior to $D1_b$, the Morar and Glenfinnan divisions were probably at different metamorphic grades. The Glenfinnan Division was in middle to upper amphibolite facies; geothermometry suggests c. 640°C at 6.4kb (section 15.2.3(i)), which would lie near the kyanite/sillimanite boundary. This probably relates to a pre-Caledonian event (cf. the other Glenfinnan Division areas). At least some Morar Division garnet is pre-$D2_b$ (Plate 9.2(f)). This gives a P,T estimate of c. 615°C at 7kb, in the kyanite field (section 15.2.2(v)). If this garnet is $MP1_b$, then it should provide a good estimate of the conditions of Caledonian metamorphism here. If it is pre-$D1_b$, then a pre-Caledonian history is probably implied for the Morar Division of this area, contrary to the views expressed by Soper & Brown (1971).

9.4.1 Slides

At several horizons within the Morar Division of this area, platy rocks similar to those found at the Sgurr Beag and other slides occur (cf. section 5.2.1; Rathbone 1980, Rathbone & Harris

1979). Also, each Lewisian inlier has a platy margin, and platy psammites extend for 50-100m into the Moines. This platiness is folded by $F2_b$ on a mesoscopic scale, and the platy zones are themselves folded on a large scale by $F2_b$ structures. It seems likely that at least two slides are present in the Morar Division of this area (cf. Fig. 9.3); Mendum (1979) has suggested that the Morar Division in Sutherland is cut up by numerous intersecting slides. The main scarp face is occupied by platy rocks; although it is also an area of high $D2_b$ deformation, the bulk of the strain appears to be of $D1_b$ age. The presence of migmatitic Glenfinnan Division rocks at the top of this zone suggests correlation with the Sgurr Beag Slide; certainly the extremely thick platy zone implies that this is the major slide in the area.

9.4.2 Lewisian

The presence of a zone of high deformation at the margins of Lewisian inliers has been mentioned above; this would have eliminated any evidence that the Lewisian was autochthonous. Although some inliers occupy $F2_b$ fold cores, the juxtaposition of Moine and Lewisian took place during $D1_b$, and there is no evidence to suggest that they occupy $F1_b$ fold cores. The apparent extension of platy rocks, at the same horizon, into the Morar Division, suggests that some factor other than rheological contrast is responsible for the high strain. The most likely explanation is that the Lewisian lies in slide zones, probably analogous to thrusts. The thickness of the Loch Naver inlier seems too large for a slice lying within a slide zone, and it may represent autochthonous basement to the overlying Moines, or a basement slice with nappe or duplex status itself. The Lewisian inlier in the Glenfinnan Division was not studied in detail, but since Glenfinnan Division lithologies are not repeated across it, it is unlikely to be a fold core, and may rest on a higher slide.

9.4.3 Migmatites

The regional migmatites of the Glenfinnan Division are distinguished here from the late granite injection of Ben

Klibreck. They contain lenticular quartz-oligoclase lits, often with biotite-muscovite-garnet schlieren, which are deformed by $F1_b$ and $F2_b$, although no obvious strain features have survived $MP2_b$ recrystallisation. The degree of migmatisation varies, with coarse semipelitic gneisses, lacking in distinct veins, becoming common on the upper slopes. Microcline is only present in lits if the host also contains it, while muscovite is common in melanosomes, making a partial melting origin unlikely (cf. section 17.1). Plagioclase compositions in lit and host are identical - although some of the Cnoc Sgriodain rocks have albitic lits (c. An_5), their host rocks, and nearby non-migmatitic pelites, show similar albitic plagioclase; this must be a function of the original sedimentary composition. The Morar Division shows no early migmatisation.

9.4.4 Granites and pegmatites

A range of variably deformed granites and pegmatites is present in this area. In the Glenfinnan Division, they can be distinguished from the regional migmatites by their sharp, often cross-cutting contacts, presence of conspicuous K-feldspar, and, in the case of pegmatites, common graphic textures.

The earliest intrusions are the granitic sills in the main scarp, and those west of the Allt a' Mhuilinn. These are concordant, medium-grained (1-2mm), and range from 10cm to 30m in thickness. The thinner bands carry an LS fabric parallel to $S2_b$, although they were never seen to be folded by $F2_b$. They are probably late $D2_b$. The large sill is only weakly foliated in the centre, and is a granodiorite with subhedral plagioclase (An_{20}), a few irregular microclines and 20-30% quartz (56619). Blue-green hornblende and some biotite is also present, along with accessory sphene and zircon. The thin sills (56620) have more granitic compositions; microcline is more abundant than plagioclase (An_{10}), and hornblende is absent. Quartz is elongated into ribbons, and feldspar twins are deformed; the lineation plunges towards 155°, parallel to $L2_b$.

The next set consists of slightly discordant granitic and pegmatitic veins, usually with north-dipping contacts (e.g. 25/005). They carry a fabric parallel to $S2_b$, although it is weaker than that in the sills. They are common from Loch na Glas-Choille eastwards, and range from a few centimetres to a few metres in thickness. Large bodies have a granitic core, and pegmatitic margins and apophyses. An example is shown in Plate 9.3(a), [565306]. The pegmatite has quartz, microcline and albitic feldspar ($<An_{10}$) in equal proportions; plagioclase, and to a lesser extent microcline, form large (few centimetres) porphyroclasts, internally unstrained, which are wrapped by recrystallised, 0.3cm polygonal matrix quartz (56612). In some rocks, feldspar is also recrystallised (e.g. 56609). The granite shows similar textures, but is finer-grained (few millimetres), and contains minor biotite; the presence of tabular plagioclase suggests that the finer grain size is original, and not due to higher deformation (e.g. 56613). Apophyses of pegmatite extend into the country rock - one lies parallel to an $F2_b$ axial surface, and shows pinch and swell structures (Plate 9.3(b)), while another penetrates at a high angle to $S2_b$, and is ptygmatically folded (Plate 9.3(c)). They thus postdate $F2_b$ folding, but have suffered some shortening perpendicular to $S2_b$. Many of the folded veins in the main scarp are probably of this set; they have S-profiles (cf. the Z-folds of banding), suggesting emplacement as a north-dipping body (Plate 9.3(d)). In the higher parts of the scarp, coarse pegmatites showing little internal deformation cut across the platy fabric (Plate 9.1(h)). Their irregular form may be due to the more heterogeneous nature of the pelitic lithologies. These veins are coarse (several centimetres), often with graphic textures, and are rich in plagioclase, which is virtually pure albite (e.g. 56618); microcline is also present. Strain is very low, with only slight undulose extinction is quartz.

All of these bodies occur as persistent veins with distinct boundaries, have compositions which are independent of their host rocks (e.g. the example in Plate 9.3(a) is emplaced in K-feldspar - free hornblendic gneiss with An_{35} plagioclase), and lack mafic selvedges. Their compositions, ranging from granodiorite to granite pegmatite, and their textures, are typically igneous, and quite distinct from the quartz-oligoclase lits of the regional migmatites. They could all be related by a differentiation sequence, with fractionation of plagioclase, and, in the main granodiorite, of hornblende and sphene (all the rocks have euhedral plagioclase *in an irregular,* quartz-microcline matrix). Plagioclase becomes more sodic in the series main granodiorite - thin sills - N-dipping granites - late undeformed pegmatites, coupled with an increase in modal microcline (although little microcline was seen in (56618), one sàmple would not be representative of so coarse a rock). This is also a time sequence, judging from the degree of $D2_b$ deformation. Successive phases have extended further west, except for the final pegmatites (which may have exploited the slide zone), possibly implying a source body lying to the east. I-type characteristics, such as presence of hornblende and sphene, and absence of primary muscovite, even in the most acid members, suggest an ultimate derivation from the mantle, or possibly Lewisian basement, rather than by partial melting of Moine rocks.

In many respects, the earlier members are comparable to the Vagastie suite; certainly the main sill is very similar to the granite at Vagastie Bridge, and except for the absence of large porphyroclasts, the petrography is identical (cf. Read 1931, pp. 85-86). (56605), one of the north-dipping pegmatites, has K-feldspar and plagioclase augen showing what Read regarded as a typical feature of the Vagastie suite, viz. coarsely perthitic microclines with the plagioclase showing albite twins in optical continuity (the larger plagioclase grains are rod antiperthites). The sills on Ben Klibreck are more deformed, and lack xenoliths or obvious cross-cutting relationships, due to the general eastwards increase in strain (section 9.2.1). The Vagastie suite is generally considered to be pre-D2 (Soper 1971). However, the main evidence is the presence of rodding parallel to L2, which does not prove that they predate the F2 folds. From Read's (1931, p. 141) description, the Loch Coire granite appears very similar to the Vagastie suite and the Ben Klibreck sheets. The simplest explanation is that all these minor intrusions are related to the granite, or to a larger phase of late-$D2_b$ intrusion of which it is a part (as distinct from the pre-tectonic Carn Chuinneag granite - cf. discussion of radiometric dates in section 9.5.2).

9.5 Discussion

9.5.1 Regional correlations

Despite its northerly position, the overall tectonic and metamorphic history at Ben Klibreck is similar to that in other parts of the Moines (Chapters 5 to 8). Glenfinnan Division rocks, which had previously suffered high grade metamorphism and migmatisation, were emplaced over non-migmatised Morar Division by a slide. This was followed by tight folding on NE-SW axial planes, with subsidiary high strain zones developed on $F2_b$ limbs. Metamorphism and recrystallisation outlasted $D2_b$, probably reaching middle amphibolite facies. An apparent difference here lies in the presence of lower slides, on which Lewisian inliers were emplaced. However, they may find a counterpart in the rings of Lewisianoid surrounding Sguman Coinntich (section 8.2.2), and in the platy zones of the Morar Division of Ardnamurchan, which are probably in part synchronous with the Sgurr Beag Slide (O'Brien 1981). The prominence here of Lewisian inliers may be due to the relatively shallow depth to basement (cf. the LISPB profile, Bamford et al. 1977, and the presence of autochthonous Lewisian at Strathy, Rathbone & Harris 1979).

Correlations thoughout Sutherland are generally based on the D2 structures with their prominent lineation. There is no doubt that $D2_b$ is the D2 of Soper & Brown (1971), at least in this area. Most workers have been prepared to trace this west to the Moine Thrust belt (Soper 1971, Soper & Wilkinson 1975), and south to the area of the Carn Chuinneag granite (Shepherd 1973, Wilson 1975). Correlation from one slide-bounded unit into another may be invalid (cf. Mendum 1979); however, if the conclusion that sliding was essentially $D1_b$ can be generalised, then correlation of D2 may be permissible, although the possibility of gradual diachronism still exists.

Identification of the slide fabric as representing a distinct event is at variance with some previous work. Mendum (1979) identified sliding as being syn-D2 in the Morar Division, while Wilson (1975) suggested a D2 age in the Garve area. It could be argued that refolded platy fabrics are the products of progressive deformation (e.g. in a shear zone), rather than distinct events. However, when the earlier fold or fabric occurs in a weakly deformed zone (e.g. the crenulation cleavage of Plate 9.2(b) or the refold in Plate 9.1(e)), this argument becomes less likely. Also, when the platy fabric is folded round large-amplitude later structures, such that the major fold does not lie entirely in a high strain zone, two phases of deformation must be represented, at least in the sense normally applied in structural analysis. Thus the presence of mappable folds of a platy Moine/Lewisian contact, on a larger scale than any $D2_b$ high strain zones, proves that emplacement of the Lewisian can be regarded as a distinct tectonic event. A similar argument can be applied to the main (Sgurr Beag?) slide, if the Cnoc Sgriodain pelites are regarded as a fold outlier; otherwise, it could be argued that only one phase of deformation postdates Glenfinnan Division migmatisation and garnet growth. Conversely, however, if the slide in a given area lay entirely within a large-scale F2 fold limb or high strain zone, it might not be possible to distinguish it as a separate event. Thus Wilson (1975) could only conclude that sliding postdated the MP1 metamorphism, predated the MP2 metamorphism, and ceased before the close of D2; similar conclusions were reached by Tanner (1971) at Kinlochourn. The model developed in areas where mappable folds of the slide are present (Mull, Acharacle, Sguman Coinntich), whereby a narrow zone of high strain is refolded by virtually coaxial and coplanar tight folds, and a weak syn-sliding fabric outside the slide zone is masked by the strong later foliation, is equally applicable in Sutherland. $D1_b$ in this thesis cannot therefore be directly correlated with D1 of other work. In the Glenfinnan Division, Wilson's (1975) D1 corresponds to the pre-$D1_b$

metamorphism; mapping further into the Glenfinnan Division would probably have revealed structures associated with this event. In the Morar Division, $S1_b$ was only identified with certainty in slide zones. The S1 fabric which is crenulated in (56001) could be of this age, or could equally be a pre-sliding fabric, analogous to that in the Glenfinnan Division. Such a situation probably obtains at Carn Chuinneag (Shepherd 1973, Wilson 1975), where F1 in the Morar Division predates at c. 560Ma Carn Chuinneag granite.

Soper (1971) and Soper & Wilkinson (1975) claim that S1 dies out upwards into the Moine Nappe, but do admit that some F1 folds deform an earlier fabric. Thus it cannot be said with certainty that no pre-Caledonian fabrics are present above the Moine Thrust, and even if this is so, such a statement cannot be extended east of the first slide in the Morar Division. The correlation by Soper & Brown (1971) of S1 in Strath Vagastie with the mylonite foliation is invalid, since all they can prove is that both are pre-D2. Since the slide fabric usually develops by intensification and attenuation of earlier fabrics, it is only readily identified when a significant drop in metamorphic grade occurs, as in the Glenfinnan Division. Thus the Morar Division of this area, and indeed throughout much of the Morar Nappe, could have a pre-Caledonian tectonometamorphic history.

9.5.2 Age of the Strath Vagastie intrusions

The age of the Strath Vagastie intrusions has been the subject of some dispute. Their pre-D2 age has been based mainly on the presence of an LS fabric, parallelling S2 in the country rocks (Soper 1971). Fergusson (1978) depicted tight folds cut by the Vagastie Bridge granite; however, these could be of F1 age. Pidgeon & Aftalion (1978) reported a 405±11Ma U-Pb zircon lower intercept, and an essentially concordant sphene result. This conflicts with 435±10Ma dates on post-D2 pegmatites at Glenfinnan (van Breemen et al. 1974), and a 430Ma date on the Loch Borrolan syenite (van Breemen et al. 1979a), which is approximately synchronous with brittle movement on the Ben More Thrust (probably

D4 in the usual Sutherland structural sequence - cf. Soper & Wilkinson 1975). Clearly if the Vagastie suite is related to the granitic sills of Ben Klibreck (section 9.4.4), then it could postdate F2 folding, insofar as a ductile fabric, parallel to S2 and L2, can develop in rocks which have already been folded, and have undergone more-or-less homogeneous shortening. This would preclude correlation of the Vagastie suite with the Carn Chuinneag granite.

Shape fabrics in quartzofeldspathic rocks are probably poor markers to use for structural correlation in areas such as overthrust terrains, where relative constancy of stress orientation over lengthy periods is likely. If a D2 shape fabric had a near-coplanar D3 strain superimposed on it, it would deform homogeneously to give a new shape, indistinguishable from the product of one phase of deformation (cf. Grocott 1979). It has been argued (section 9.2.1) that a lengthy period of homogeneous straining postdated $F2_b$ folding, and lasted through the period of granite and pegmatite emplacement. Although this has been included in $D2_b$, the later increments could have developed during D3 or D4 of other areas: the change to an ESE extension direction in the latest veins suggests some long-term variation in stress orientation (note that despite this, no refolding of the 155°-trending lineation is seen in adjacent psammites).

9.5.3 Precambrian events

The possibility of a Precambrian event in the Morar Division makes the relations of garnets particularly critical. Soper & Brown (1971), Shepherd (1973) and Wilson (1975) have all claimed that regional garnet is MP2 in age. The last two noted MP1 garnet in the aureole of the Carn Chuinneag granite, and (citing its absence outside the aureole) suggested that it was due to contact metamorphism, although suitable pelite horizons are scarce elsewhere. Soper & Brown also noted kyanite and staurolite in the Loch Meadie schists, to the west of Altnaharra, and used their orientation parallel to the L2 lineation to infer an MS2 age.

However, porphyroblasts could have been rotated by D2 homogeneous strain. There is little doubt that the MS2-MP2 metamorphism in this area exceeded garnet grade; the question is whether any significant earlier (i.e. Precambrian) event is represented. Given the structural arguments above, Precambrian deformation cannot be ruled out, and at least some garnets here are pre-$D2_b$ (Plate 9.2(f)). Geothermometry applied to such garnets (section 15.2.2(v)) gives a result (615°C at 7kb) which is consistent with the southern Morar Division where a Precambrian metamorphism is established. It also bears the expected relationship (lower temperature at broadly similar pressure) to the Ben Klibreck Glenfinnan Division (cf. section 15.3.2). The pressure is rather high for an area only 30km from the foreland, and might be more easily accommodated in an earlier, Precambrian metamorphism. The high pressures contrast with the presence of sillimanite on Ben Klibreck (Read 1931). This may be Caledonian in age, related to the granite intrusions in the area; certainly that shown by Read (1931, figs. 10,11) appears to be late (cf. also Watson 1948).

9.5.4 Origin of migmatites

Many of the anomalous features in the migmatites of Ben Klibreck are due to the operation of two distinct processes. As in other areas, the oligoclase-quartz lits of the Glenfinnan Division developed by segregation during high grade Precambrian metamorphism (sections 9.4.3, 14.1), probably by isochemical, sub-solidus processes (sections 15.3.1, 16.2, 17.1). The granitic veins and late pegmatites form part of a late Caledonian igneous suite, possibly related to the Vagastie suite and the Loch Coire granite, and are essentially intrusive. The scattered chemical results of Brown (1967) are due to indiscriminate sampling of both sets. The argument that sodic pelites corresponding to the regional migmatites are absent in the Morar Division is irrelevant, since the two divisions represent entirely different structural/ stratigraphic units (cf. Johnstone 1975). The order of emplacement of the granites is essentially that of Soper & Brown (1971), except for their sodic pegmatites, which appear to correspond to the regional migmatitic lits. These authors place the western boundary of the migmatites at the Loch Naver inlier. In this thesis, two distinct boundaries have been identified - the slide underlying the Glenfinnan Division regional migmatites, and a more diffuse boundary to granite intrusion which extends well into their "zone of veins"; these also extend eastwards into the Glenfinnan Division. The first boundary coincides with that between Read's (1931) zone of veins and zone of injection.

CHAPTER 10

Lochan Coire Shubh

10.1 Introduction

An area south of Lochan Coire Shubh, 2km SSE of Kinlochourn, was chosen for a study of the structural relationships of the later (Caledonian) pegmatites. The country rocks lie entirely within the Glenfinnan Division, some one to two kilometres east of the Sgurr Beag Slide (cf. Chapter 5).

Two lithostratigraphic units can be mapped (Fig. 10.1). One consists dominantly of highly garnetiferous migmatitic pelite, with very minor semipelite and psammite, although a thick psammitic bed occurs 100m from its boundary. The other consists mainly of somewhat feldspathic psammite, with minor quartzite and semipelite, and appears to define a major fold interference pattern, in the core of which the pegmatite complex described below is centred. Individual lithologies are similar to those at Kinlochourn (section 5.1); no Lewisianoid rocks or garnetiferous amphibolites were seen, but calc-silicates are widespread in the pelite.

Regional metamorphic grade (prior to development of the pegmatite complex) was probably kyanite-sillimanite, as at Kinlochourn (section 5.3).

10.2 Structure

Two major phases of folding can be recognised: tight-to-isoclinal folds (which are at least F2 in age), and open, NE-SW folds, which define the main mappable structure. They are conveniently described in reverse order.

10.2.1 $F3_1$

The map of bedding-foliation (Fig. 10.1) shows a major, upright, SW-plunging antiform, with axial plane following the Allt Coire Shubh Beag. Related minor structures are open-to-close, upright folds on axial planes trending 020° to 045° (Plate 10.1(a)); they lack an axial planar cleavage, but a weak shape fabric is developed in some psammites. Open crenulations of suitably orientated earlier fabrics are common, and open folds (Plate 10.1(b)), often associated with a weak cleavage, affect some of the late pegmatites. Most of the late pegmatites carry a weak NNE fabric, and can probably be regarded as late $D3_1$ in age; saddle reefs in the major fold are filled by pegmatite.

10.2.2 $D2_1$

Numerous examples occur of tight-to-isoclinal folds of bedding and migmatitic banding, which exhibit a strong axial planar cleavage (Plate 10.1(c)). They are overprinted by weak $F3_1$ crenulations, and cut by the late pegmatites (Plate 10.1(d)). Their axial planes trend sub-parallel to the bedding-foliation (Plate 10.1(c)), and plunges are generally steep to the west or northwest. The axial planar fabric is a crenulation cleavage of an earlier schistosity, and deforms the regional migmatitic lits (Plate 10.2(a)), and hence will be termed $S2_1$. Alternate folded and platy zones are crossed, in traversing at right angles to the bedding-foliation; in the platy zones, early migmatitic lits are intensely deformed (Plate 10.2(b)). No systematic distribution of S *versus* Z profile $F2_1$ folds could be mapped; however, the centre of the psammite outcrop marks a zone of abundant M and W $F2_1$ folds, contrasting with the alternating platy zones, and S and Z folds, of the outer and inner arcs. This suggests that an $F2_1$ fold core does lie within the psammite outcrop, and that the major structure is part of an interference pattern (Type 2 or 3 of Ramsay 1967).

10.2.3 Discussion

The structural sequence in this area is broadly similar to that found elsewhere in the Glenfinnan Division (e.g. sections 5.2, 6.2, 11.2). The closest comparison can be made with Kinlochourn (section 5.2), whose $D2_k$ structures are very similar to $D2_1$. $D3_k$ structures are generally tighter than $D3_1$, but their correlation is favoured by the abundance of syn- to post-$D3_k$ pegmatites at Kinlochourn, and the late-$D3_1$ age of the pegmatite complex at Lochan Coire Shubh.

10.3 Pegmatite complex

The area south of Lochan Coire Shubh, particularly the core of the major fold, is intensely veined by pegmatites (often making 60-70% by volume). These are coarse, graphic-textured, and consist of quartz, oligoclase and K-feldspar. The earliest pegmatites are very coarse-grained (several centimetres), mica-poor, and form concordant lenses and sheets in the core of the fold at [958044]. These have sharp contacts, contain disoriented xenoliths, and presumably were intruded into voids opening up in the crest of the antiform. Similar pods occur in the cores of minor antiforms, and carry a moderately strong vertical shape fabric, trending 030° (Plate 10.2(c), (d)). The next intrusions are most common in the area around [958046], and form schollen migmatites in SW-dipping psammites; these also carry the 030°-trending foliation. They are cut by a major swarm of broad, NE-SW dykes, with NW-SE apophyses (Fig. 10.1). These are often composite, and show marked reaction zones against the host psammites (Plate 10.3(a)). The zones are relatively enriched in mica and garnet, and strongly depleted in feldspar, and are probably restites, i.e. psammitic material from which a pegmatite composition has been extracted. All the pegamatite in the dyke swarm could not have been obtained from the local (relatively feldspar-poor) psammites, but the abundance of similar irregular restite patches, even in dykes showing no reaction zones, suggests derivation from similar rocks at depth. In places, the proportions of pegmatite/restite/psammite are consistent with local derivation, and even with expulsion of pegmatite from the immediate area (Plate 10.3(b)). Many of these veins appear to be intrusive, result in minor misorientation of psammite blocks, and show flow-like patterns of muscovite books. Both the locally produced, and the introduced material, was probably molten. Elsewhere (e.g. [961043], Plate 10.3(c)), quartzites are agmatised by quartz-rich aplites which carry the 030° fabric; these were probably produced by localised partial melting at about the same time as the dykes. All the above bodies are cut by a second major swarm of coarse, muscovite-rich NW-SE pegmatites. They form individual dykes, zoned towards their centres, and preserve the regional orientation of banding in septa of country rock (Plates 10.1(d), 10.3(d)), although they still include rare patches of restite. The dyke swarm defines an open fold, whose axial plane coincides with that of the major $F3_1$ antiform (Fig. 10.1). It may have been intruded as a linear swarm before the $F3_1$ fold had attained its present amplitude (i.e. syn-$F3_1$). Some of the pegmatites carry a weak 030° fabric, continuous with $F3_1$ crenulations, and their margins show open folds on NE-SW axial planes (Plate 10.1(b)). Minor folds of bedding and pegmatite are not always congruent, indicating that the pegmatites were emplaced fairly late in the $D3_1$ event, e.g. during relatively homogeneous shortening after the folds had formed.

10.4 Discussion

At Lochan Coire Shubh, the Glenfinnan Division regional migmatites have been structurally reworked in a similar fashion to their counterparts at Kinlochourn and elsewhere. However, the "late migmatisation" of sections 5.4.2 and 6.4.2 is much more intense, and leucosomes consist of both locally-derived and introduced material. The widespread partial melting or segregation of country rock occurred during and immediately after the climax of pegmatite intrusion, and may be related to their thermal effect, although if the pegmatites themselves were derived from only slightly greater depth, a local increase in heat flow may be ultimately responsible. The localisation of the complex in the core of an $F3_1$ antiform may be due to selective intrusion in the fractured psammites of the axial region, or channelling of heat or fluids from deeper dehydration reactions.

The later dykes are identical to the late- to post-$D3_k$ intrusive pegmatites at Kinlochourn; one of these (LH23, Appendix 3) was analysed. Its composition (norm Quartz 32%, Orthoclase 29%, Plagioclase (An_{20}) 33%) and rare earth pattern is consistent with derivation by a large degree of partial melting of Moine psammites or semipelites, with consumption of virtually all the available plagioclase, leaving a quartz-mica-garnet restite (section 16.3.4). Patches of restite are generally absent from these pegmatites, suggesting that they have travelled some distance from their source, which may be the Lochan Coire Shubh complex, or some other local hot spot.

CHAPTER 11

Loch Quoich

11.1 Introduction

The area at the head of Loch Quoich is of interest for the presence of the Quoich granitic gneiss, and of the Loch Quoich Line, which separates the Glenfinnan and Loch Eil divisions of the Moine (Chapter 2; Johnstone 1975). The Quoich granitic gneiss is one of a series of bodies which run approximately N-S, parallel to and in the region of the Glenfinnan/Loch Eil boundary (Figs. 11.1, 17.1), before turning east just north of here and trending towards the Great Glen at Fort Augustus. These have a uniform adamellitic composition (e.g. Gould 1965), and have been variably interpreted as metasomatised semipelites (Dalziel 1966), slices of basement (Harris in discussion of Winchester 1974), or pre-metamorphic granites (Pidgeon & Aftalion 1978). A section through the gneiss in the spillway of the Quoich dam was chosen for study, as well as country rocks in the quarry and island in Loch Quoich (Fig. 11.1). No large-scale lithological map was made; however, a 1:1000 plan of the spillway is shown in Fig. 11.2. Geochemical arguments for the mode of origin of the gneiss will be presented in sections 16.3.3 and 17.3.1; however, it will be shown below that it was in its present state prior to the first regionally correlatable event in the country rocks, timing which is incompatible with Dalziel's model. The main lithologies observed are described below.

Granitic gneiss In this area, the granitic gneiss is about 70m thick, and is essentially parallel to banding in the meta-sediments, although some boudinage of a marginal pelitic band occurs at the western contact. It is a medium-grained (2mm) granular rock, consisting of approximately equal amounts of quartz, plagioclase (An_{15}) and K-feldspar (norm Q 38 Or 30 Plag 27), with 3-5% biotite and minor garnet, apatite, epidote and zircon - e.g. (56173). The opaque phase is usually pyrite and/or ilmenite, but haematite is often present as an alteration product. The rock is banded on a centimetre scale, by segregation of orientated biotite flakes into foliae which define the gneissic fabric; thicker pegmatitic or aplitic veins and lenses lie sub-parallel to this foliation and lack mafic minerals. The gneiss is homogeneous on the scale of a few metres, but variations, particularly in biotite content, occur across the section. It has a sharp western boundary, but the eastern one is more transitional, in the sense that it is interbanded with the metasediments, rather than having margins of intermediate composition. A number of enclaves of metasediment (quartzite, hornblende schist and biotite pelite) are present near the eastern boundary, the largest defining a fold interference pattern (Fig. 11.2). These are probably best regarded as highly deformed xenoliths or roof pendants. An angular xenolith occurs in less deformed granitic gneiss to the south of the loch (A.M. Roberts, verb. comm.). The gneiss has suffered local mobilisation, to the extent that it agmatises early basic dykes (Plate 11.1(a)).

Metasediments The metasediments in the spillway are dominantly psammitic, although several thick migmatitic pelites occur downstream (Fig. 11.2). The island in Loch Quoich consists mainly of coarse, garnetiferous, lit-par-lit migmatitic pelite, carrying abundant lenses of garnetiferous amphibolite, while the quarry contains flaggy, feldspathic psammites. The contact between dominantly pelitic and dominantly psammitic rocks occurs by the roadside at approximately [064028] , and has been regarded by I.G.S. as the Glenfinnan/Loch Eil division boundary. Calc-silicates are present throughout, including both the typical garnet+hornblende-bearing types, and others, containing abundant epidote but lacking mafics, which recall Ramsay & Spring's (1962) Arnipol type. All these lithologies are typical of the Glenfinnan Division, and the dominantly psammitic Loch Eil Division is distinguished only by the proportion in which they are present; there is no apparent metamorphic or tectonic contrast between the divisions which would suggest an unconformable relationship (cf. Lambert et al. 1979).

Pre-metamorphic igneous rocks Two types of early amphibolite are present. Coarsely garnetiferous amphibolites are typical of the Glenfinnan Division, and consist mainly of garnet and hornblende, with minor andesine and quartz. In Mull, similar rocks are clearly cross-cutting (Plate 7.2(a)), but here they are thin and lenticular, and are generally considered to represent metasediments or tuffs (e.g. Johnstone *et al.* 1969). In the Loch Eil Division, stripes and lenses of hornblende-biotite (± garnet) schist (e.g. 54677) occur, and are probably analogous to those in the Glenfinnan Division. Both these types predate $D1_q$ of this chapter, although in view of its polyphase nature (section 11.4) this need not imply an entirely pre-tectonic origin. A suite of coarse amphibolite dykes is present in the granitic gneiss and in its enclosing metasediments; these are sub-concordant, range up to 1m in thickness, and consist typically of 75% hornblende and 25% An_{40} plagioclase. They postdate the bulk of $D2_q$ strain, but are deformed by the shear zones, and preserve no igneous textures.

Late- to post-metamorphic igneous rocks A variety of late, K-feldspar-rich pegmatites are present. The earliest are concordant and biotite-bearing, and frequently show pinch-and-swell structures. Some are emplaced along $F3_q$ axial planes, although their essentially undeformed state shows that they must postdate the bulk of $D3_q$ strain. Later, cross-cutting pegmatites fall into two groups; a 175°-trending set of two-mica pegmatites, in which feldspar exhibits undulose extinction and quartz is granulitised, and a vertical, 015°-trending, muscovite-rich set, in which quartz is only slightly granulitised (e.g. 54669). These are cut by thick microdiorite dykes, which generally lie in moderately SE-dipping fault planes, and have a foliated margin indicative of westerly overthrusting (foliation steeper than contact). Textures are essentially metamorphic, although relict ophitic texture occurs in the cores of some thicker dykes. Felsic porphyrites, possibly belonging to the same suite (Smith 1979), also occur, and appear to be somewhat later, insofar as they are more strongly cross-cutting (e.g. some trend E-W with irregular boundaries) and less strongly foliated. Large, zoned plagioclase phenocrysts are preserved (Plate 11.1(b)), but the matrix is recrystallised to a polygonal texture. Several late hornblende-plagioclase intrusions occur in the upper spillway, with irregular contacts showing net-veining and brecciation of psammites, and a weak, margin-parallel foliation.

11.2 Structure

A sequence of mesoscopic structures was defined in two small areas (spillway and island/quarry); their relationship to the regional geology can be seen on Fig. 11.1 based on I.G.S. 6-inch sheets.

11.2.1 Spillway

The full sequence of structures is best seen in the granitic gneiss, although all seem to have counterparts in the metasediments. Their main features are summarised below.

$\underline{F1}_q$ $S1_q$ is the earliest foliation in the granitic gneiss; it is only clearly seen in $F2_q$ fold cores, where a segregated gneissic fabric, similar to $S2_q$, is preserved (Plate 11.2(a)). A foliation runs parallel to the margin of isoclinally-folded lits in the migmatitic pelites, and pre-$F2_q$ isoclinal folds of bedding are present in the large xenolith, near the eastern margin of the granitic gneiss (Fig. 11.2(d)).

$\underline{F2}_q$ This event gave rise to the dominant foliation in gneiss and metasediment. In the granitic gneiss it is a planar, segregated fabric, with biotite seams a centimetre or so apart. It is axial planar to tight (interlimb angle < 10°) folds of pegmatitic lits (Plate 11.2(b)), and to folds of $S1_q$ in the main xenolith. The amphibolite dykes cut this foliation, and carry xenoliths which contain folds, but where slightly cross-cutting, they are weakly foliated and folded (Fig. 11.2(b)), probably as a response to relatively homogeneous shortening within the gneiss. In the metasediments, tight folds with an axial planar foliation are probably of this generation; these are particularly common in the quartzites west of the granitic gneiss. These folds plunge steeply southwards, and in the western part of the spillway, indicate an antiform to the east (Fig. 11.2(a)). Plunges steepen eastwards to become vertical, and within the granitic gneiss, $F2_q$ folds of the gneissic foliation, and of the xenolith, plunge northwards, indicating an antiform to the west. However, since S-profiles are maintained throughout when viewed down-plunge, a fold core is probably not present - i.e. the fold axes have passed through the vertical, and lie on the same limb of a larger fold (the core of which was not identified in this area).

The $S2_q$ fabric is deformed by vertical shear zones, which are particularly common in the eastern part of the granitic gneiss, although they are also present in the metasediments. The zones are generally a few centimetres wide, and a few metres long, with tens of centimetres displacement (Plate 11.2(c)). Their profiles appear consistent with simple shear, but as they are usually the sites of preferential segregation or melting, no detailed measurements were made. Sinistral members are by far the more common, trending 145°±5, but several dextral examples occur, their trend of 060° implying an internal angle of about 85°, with compression from 110-115° (orientation based on sense of observed displacement, not just acute angle). These zones deform and displace the thinner basic dykes, with development of a new schistosity. Pinch-and-swell in thicker dykes is filled by pegmatitic material derived from the shear zones, and the folds in Fig. 11.2(b) may date from the same period of WNW-ESE shortening. Small-scale conjugate folds indicating layer-parallel shortening are particularly common near the margins of the granitic gneiss (Plate 11.1(c)). These are always vertically plunging, and dextral members with NW-SE axial planes are the more common. They often show strong recrystallisation and mobilisation in their cores, possibly with concentration of partial melt (Plate 11.1(d)), apparently giving rise to a large local volume increase in the cores of the folds at the eastern boundary of the granitic gneiss (Fig. 11.2). At one locality (Fig. 11.2(e)), shear zones are reorientated by such a fold, confirming the relative order of the two events. None of these structures has any large scale expression, or gave rise to large bulk strains (e.g. shortening of the granitic gneiss by the shear zones would only add up to a few percent). Their association with segregation, and possibly partial melting, indicates thermal continuity with $D2_q$ rather than $D3_q$.

<u>$F3_q$</u> This event gave rise to large-scale, open-to-tight structures of variable orientation. In the spillway, minor fold pairs of about one metre wavelength, lacking an axial planar foliation, indicate an antiform towards the west. Axial planes dip at about 60/255 with plunge of axes ranging from 25/180 to 45/205. At the head of the spillway, and in the road cuttings (Fig. 11.1), a large, S-plunging antiform with upright, N-S axial plane occurs, and further west, a series of upright fold pairs was seen, with Z-profiles looking north.

<u>Later structures</u> Open, ductile folds with sub-horizontal axial planes, some of which may be related to $F2_q$, occur throughout the steep rocks. A number of large kink folds, which pass into vertical faults displacing amphibolite dykes, were recorded (Fig. 11.2), and the spillway itself follows a zone of vertical fractures, some showing small displacements, which are associated with reddish staining.

11.2 Island and quarry

Structures in this area are best displayed in the pelites exposed along the shore of the island at [063017]; the psammites of the quarry are dominated by strong fabrics sub-parallel to bedding. The sequence is directly comparable to that in the spillway area.

<u>$F1_q$</u> Evidence for this generation of structure is seen in interference patterns on the southeast shore of the island. A schistosity which contains the migmatitic banding is folded by $F2_q$ (Plate 11.2(d)), and this fabric contains intrafolial folds in some psammites. The foliation in garnet amphibolites is also of $F1_q$ age.

$\underline{F2}_q$ In the quarry, a strong foliation, trending about 5° clockwise of bedding, and an associated intersection lineation are probably of this age (Fig. 11.3(b)). Intrafolial folds showing the same vergence were seen in pelites (Plate 11.1(g)). On the island, a number of small- to medium-scale, reclined $F2_q$ folds occur, including one with >10m wavelength which can be traced for >100m (Fig. 11.1). It lies on one limb of a gently-plunging $F3_q$ fold, and carries an axial planar foliation which is absent from the latter. Shear zones and conjugate folds, which are similar to those in the granitic gneiss, are present although relatively uncommon (Plate 11.1(e), (f)). The conjugate folds are difficult to distinguish in areas of steeply plunging $F3_q$.

$\underline{F3}_q$ This phase is represented by a number of tight, upright, NNE-trending folds, usually lacking an axial planar cleavage (Plate 11.1(h)). Plunge of axes is highly variable, ranging from about 10/200 in the south of the island, through 25/020 in the north, to 50/010 in the quarry (Fig. 11.3(b)), and curving through 25-30° in individual exposures. These folds consistently verge towards an antiform to the east.

Later structures Rare kink-like folds occur, and open, flat-lying folds coaxial with $F3_q$ are common. Some at least of these are later - e.g. those in (56122) fold grain-size reduction zones, which themselves are probably post-$D3_q$.

11.2.3 Discussion

Despite the scattered nature of the exposures, a unified picture does emerge.

$F3_q$ folds are open to tight, and lack an axial planar cleavage. They have axial planes which are west-dipping in the east, and steepen towards the west, concurrent with a general tightening of the structures (this can be observed in the road section north of the dam). Opposing vergence at the quarry and the dam seems to indicate the presence of an antiform. However, $F3_q$ axes plunge in opposite directions at these localities; if they passed through the vertical, then no closure would be implied, and the structure would be analogous to the middle limb of the Sron na Gaoithe fold (section 6.2; cf. Fig. 11.3(c)). From Fig. 11.1, the major fold of the granitic gneiss is probably an $F3_q$, and the southward-closing fold north of the quarry may be its complement.

A similar situation occurs with $F2_q$ - an apparent antiform lies in the granitic gneiss (Fig. 11.2); however, $F2_q$ fold plunges are seen to pass through the vertical in the quartzites. Moving westwards, $F2_q$ axes become reclined on the island, and return to north-plunging in the quarry, although $S2_q$ consistently lies clockwise of bedding. If fold axes are continuously passing through the vertical, antiforms and synforms have no meaning; it is more useful to consider the profile of the fold on the ground, which remains constant, and to define north-closing or south-closing folds. On this basis, assuming that the $F2_q$ closure on the island is of only local significance, no major $F2_q$ or $F3_q$ structures are present in the area between the quarry and the spillway.

$F1_q$ includes all pre-$F2_q$ structures. It is likely that more than one generation is represented - cf. the correlations in section 14.3. Occasional veins in pelites, which appear to be folded by $F1_q$, may have formed between two events which now comprise $F1_q$ because they can no longer be distinguished. They would be analogous to the MP1 veins of Knoydart, which predate the main migmatisation (section 4.3).

11.3 Metamorphism

Metamorphic textures in this area are largely obliterated by high late deformation and, particularly in the spillway, intense retrogression. Textural evidence will be discussed first, followed by less direct inferences based on migmatitic features.

11.3.1 Mineral textures

Biotite-grade conditions or higher prevailed until after $D3_q$, since biotite grew in $F3_q$ crenulations. The garnet-hornblende-An_{40} plagioclase assemblage in post-$D2_q$ basic dykes indicates middle to upper amphibolite facies conditions for $MP2_q$ or $MS3_q$. Similarly high grade conditions are indicated by the garnetiferous amphibolites; however, this relates to a pre-$D2_q$, and possibly pre-$D1_q$ event (Fig. 11.3(d)). Garnets in the migmatitic pelite predate $S2_q$ at least (Fig. 15.4(e)). The microdiorites and felsic porphyrites have recrystallised to amphibolite facies assemblages; however, this may in part be due to the internal heat of the dyke, so that the country rocks may only have been in the greenchist facies (cf. Smith 1979). Some support for this is seen in the westernmost dyke on Fig. 11.2. The fine-grained margin carries hornblende and fresh andesine, while the coarser centre has a more actinolitic amphibole and saussuritised plagioclase, a pattern which could develop through slow cooling in hot country rocks. If the dyke cooled, and was then affected by a prograde metamorphism, then the margins should be more severely altered. Retrogression is so intense that much of the granitic gneiss now consists of large crystals of albite containing zoisite granules, although the rock appears fresh. All calc-silicates were also intensely retrogressed.

11.3.2 Migmatisation

The area displays an extended history of quartzofeldspathic segregation, which may generally be described as migmatisation. The pelitic gneiss, and some psammites and semipelites, are coarse lit-par-lit migmatites of typical Glenfinnan Division type - i.e. muscovite-bearing, with quartz-oligoclase lits, the latter typically measuring one centimetre by a few tens of centimetres, and probably having been formed by subsolidus processes (section 17.1). They are very early ($MP1_q$ or older), and most of the garnet in the metasediments is probably of approximately the same age. A P,T estimate of 590°C at 5.5-6kb (section 15.2.3(vii)) is comparable to those from the Knoydart Pelite, and would indicate kyanite grade.

The lits in the granitic gneiss are pre-$D2_q$; in Ardgour, similar segregations or pegmatites are of $MP1_g$ age (section 12.2), possibly correlating with migmatisation in the metasediments. They are probably also of subsolidus origin. No segregations of $D2_q$ age could be identified; although most early lits now lie in $S2_q$, this is invariably due to transposition (Plate 11.2(b)).

The next major episode of remobilisation was associated with the late $D2_q$ shear zones and conjugate folds. The shear zones show marked coarsening with depletion in mafics, and ultimately development of pegmatites, which may intrude the surrounding gneiss (Fig. 11.2(c)), and agmatise basic dykes. The larger conjugate folds have their cores replaced by granular granitic material, which appears to have been molten, and which deforms and metasomatises (by growth of large feldspar porphyroblasts) thin basic dykes. Their origin is discussed in section 17.3.2. The contrast in style of emplacement of successive cross-cutting pegmatites clearly indicates declining country-rock temperatures post-$D3_q$.

In summary, the area underwent an early, probably kyanite-grade metamorphism, followed by intense deformation, and then a further high-grade event, which may have reached partial melting temperatures. Slow cooling followed, so that all strain features were annealed, and thorough low grade recrystallisation occurred.

11.4 Conclusions

Due to the intense deformation and metamorphism, evidence for the nature and origin of the granitic gneiss is difficult to obtain. The most important conclusion is that it shows all the tectonic events recognised in its country rocks. It was a granitic gneiss by the end of $D1_q$, and has probably suffered the early high grade metamorphism which created the regional migamites. This, coupled with the absence of zones of unusually high strain at its contacts, makes an origin as a slice of basement emplaced into low grade Moines unlikely. There is no evidence to support a metasomatic origin, and its internal homogeneity, with sharply interbanded rather than gradational contacts, argues against such a hypothesis. Bearing in mind the probable xenoliths, the simplest explanation is that it was a granitic intrusion which predated the peak of metamorphism, but may have postdated some of the structures in the metasediments (full discussion in section 17.3.1).

The Loch Quoich Line passes through this area, and supposedly separates steeply-dipping, migmatitic, striped Glenfinnan Division rocks with steeply-plunging structures, from gently-dipping Loch Eil Division psammites with open, gently-plunging structures. Further south, Dalziel (1966) postulated a mobile infrastructure of Glenfinnan Division migmatites, overlain by the Loch Eil Division, with the granitic gneiss developed metasomatically at the highly strained contact between the two; Lambert et al. (1979) suggested that the Loch Eil Division may be unconformable on already-metamorphosed Glenfinnan Division. It is clear from the present work that three distinct phenonomena are represented, which are temporally and spatially independent of one another. The lithostratigraphic boundary runs west of the dam, but there is no evidence for a structural or metamorphic break at this position; it appears to be a normal stratigraphic boundary between a striped and a psammitic sequence. This boundary occurs within the steep belt, and from the I.G.S. maps, appears to be folded by at least the $F2_q$ and $F3_q$ structures. The granitic gneiss lies within the Loch Eil Division, but still in the steep belt, and to the northeast climbs to higher levels within the Loch Eil Division as it enters the flat belt; it was emplaced prior to $D1_q$. The steep belt/flat belt boundary appears to reflect the tightness of $F3_q$ folds; within the Loch Eil Division of this area, and in road cuttings as far east as Invergarry, the gently-dipping psammites are intensely deformed in a flat-lying fabric, presumably composite $S1_q$-$S2_q$. The difference in fold style is probably due to the broad lithological contrast (cf. the change in $F3_k$ fold style across the Sgurr Beag Slide at Kinlochourn - section 5.5, Fig. 5.2), and is a relatively late Caledonian feature.

As at Acharacle (section 6.5), there is no obvious explanation for the curving of fold axes through the vertical. The highest strains are associated with steep plunges, which makes models involving sub-horizontal extension implausible. Both $F2_q$ and $F3_q$ structures show this phenomenon (as opposed to $F3_a$ only at Acharacle); they appear to vary independently, on the limited evidence available (Fig. 11.3(a)). With more widespread mapping, it might prove possible to use the variation of both elements to provide better constraints on possible mechanisms (e.g. if a model involving superimposed strains was invoked, it should be possible to determine if one event could curve both sets of axes in the observed manner, or if two independent events would be required).

The structural sequence here precisely matches that at Glenfinnan (section 12.2), and bears a broad resemblance to those in more westerly areas, particularly Lochan Coire Shubh (section 10.2). Regional correlation (section 14.3) suggests that $D1_q$ represents more than one event (probably encompassing the whole pre-Caledonian history of these rocks), so the relationships of metamorphic or migmatitic textures to particular $F1_q$ structures cannot be generalised - i.e. it is best to regard the regional migmatisation and early metamorphism, and possibly intrusion of the granitic gneiss, as broadly $D1_q$ in age.

CHAPTER 12

Glenfinnan

12.1 Introduction

The new road cuttings between Glenfinnan and Kinlocheil were chosen to provide a section through the Ardgour granitic gneiss (Dalziel 1966). This essentially adamellitic body (Gould 1965) is one of a series which run approximately along the Glenfinnan Division/Loch Eil Division boundary. Their regional setting was discussed in section 11.1. Dalziel (1966) mapped the area in detail, producing a structural and stratigraphic synthesis which is summarised in Fig. 12.1; the three metasedimentary units were considered by Dalziel to be essentially in stratigraphic order, and the granitic gneiss to have been derived metasomatically from the Beinn an Tuim striped schists. The major structures are F2 and F3 in age in Dalziel's fold chronology, two of the former being lost in a zone of high strain near the western margin of the granitic gneiss.

In field appearance and thin section, the granitic gneiss is identical to that at Loch Quoich (section 11.1), except in the $F2_g$ fold core (see below). There, it is noticeably less deformed, with pegmatitic lits oblique to the gneissic foliation; it is also coarser-grained, with subhedral plagioclase more prominent (e.g. 58144). Amphibolite dykes are also present (e.g. 58147), and are cut by coarse, graphic pegmatites, which themselves are cut by foliated microdiorites (cf. section 11.1; Johnson & Dalziel 1966). The present study was concerned mainly with the mesoscopic structures in the granitic gneiss, and their comparison with those at Loch Quoich.

12.2 Description of traverse

The first exposures (at (a) on Fig. 12.1) consist of gently-dipping granitic gneiss; Loch Eil Division psammites were seen 1km to the east, but the intervening pelites are not exposed in the road section. The dominant foliation is deformed by upright folds ($F3_g$), plunging at approximately 10/040, with vergence indicating that a local antiform lies to the east. Earlier, tight-to-isoclinal folds ($F2_g$), with an axial planar foliation, have northeasterly plunges, and an overall Z-sense (Plate 12.1(a)). The next exposures occur west of Dubh Ligh (b), and consist of homogeneous granitic gneiss with 1-2cm thick pegmatitic lits. Dips vary from 40/080 to 10/320, due to open folding on gently NNE-plunging axes; some lits outline fold cores to which the $S2_g$ gneissic fabric is axial planar.

At (c), dips are still variable, but western limbs of $F3_g$ synforms steepen to a vertical, NNE attitude, indicating an antiform lying to the west. E-plunging $F2_g$ folds of lits are present, and maintain their Z-profiles and reclined orientation. At (d), foliation dips gently, with indications of an $F3_g$ antiform to the east, and the reclined $F2_g$ folds are joined by shear zones, which show a normal sense of displacement and trend east-west. Some S-profile folds may be of $F2_g$ age, or could be ptygmatic folds of post-$F2_g$ pegmatites. At [921799], a large, sill-like pegmatite disrupts basic dykes which are similar to those at Loch Quoich (section 11.1).

To the west, the gneissic foliation steepens in a series of large steps, verging towards an antiform to the west; steep limbs trend about 020°. At (e), the foliation again dips gently ESE, and is axial planar to open folds of lits and of an earlier gneissic fabric. These folds, whose axes plunge gently to 060°, show no vergence, and the weakness of $S2_g$ suggests that a major $F2_g$ fold core is present. In Plate 12.1(b), the pegmatitic lits are clearly pre-$F2_g$, and are being transposed into the $S2_g$ fabric (a late, ptygmatically-folded pegmatite is also seen). In Plate 12.1(c), an earlier, $S1_g$ gneissic fabric is preserved, which is oblique to the lits, although the original angle was probably less than the maximum now observed (cf. Ramsay 1967, fig. 9.32). This favours an $MP1_g$ origin for the lits. At [91858010], a strong gneissic fabric dips at 80/350, and is cut by steeply south-dipping pegmatites. This is probably $S1_g$, in the core of a major $F2_g$ fold.

At (f), $S2_g$ is stronger, and is axial planar to ENE-trending folds of lits; it steepens westwards through 50/065 to 80/110, due to $F3_g$ folding. By (g), $S2_g$ is penetrative, with only rare intrafolial folds preserved, and $F3_g$ folds are tighter, with more consistently vertical, NNE-trending axial planes (Plate 12.1(d)). The plunge of the later folds steepens westwards, and in addition they clearly indicate an antiform to the west.

Further west, typical Glenfinnan Division migmatitic pelites and semipelites appear, across a sharp, concordant contact. Folds similar to $F3_g$ continue to indicate than an anitform lies to the west, and have a strong crenulation cleavage; intrafolial $F2_g$ closures are reclined, and generally show S-profiles.

12.3 Discussion

An early gneissic fabric ($S1_g$) and pegmatitic lits are deformed by isoclinal folds, to which the dominant gneissic fabric is axial planar. The lits (e.g. in 58145) are probably post-$F1_g$, making an igneous origin unlikely. Comparison with the metasediments suggests that the granitic gneiss was intruded and deformed prior to a static metamorphism, which probably coincided with the regional migmatisation. Quartz and feldspar are largely recrystallised, but primary zoning is preserved in some lits, from a pyrite centre, through lath-like plagioclase, to coarse, equant quartz and plagioclase (K-feldspar may also be present). Plagioclase in lits has a composition identical to that in the host gneiss (An_{12}). The origin and structural relations of the gneiss and its lits will be discussed in detail in section 17.3.

$F2_g$ folds appear to indicate the presence of a fold core within the granitic gneiss; vergence of minor folds (S-profile in the west, Z-profile in the east) suggests a southward closure. The low $D2_g$ strain and neutral $F2_g$ vergence at (e) confirms that an $F2_g$ fold core is present, and that the opposing vergence is not produced by $D3_g$ rotation. $F2_g$ axes plunge east or ENE, although $F3_g$ rotation means that axial planes change from horizontal to vertical. If $F3_g$ was unfolded to leave $F2_g$ subvertical and NNE-trending, a steeply NNE-plunging synform would be indicated, similar to Dalziel's (1966) Meall nan Damh fold (MDF on Fig. 12.1). If it is assumed that $S2_g$ was originally flat-lying, a southward-closing, nearly reclined fold, with eastward-trending axis, would be indicated. Major lithological units are not symmetrical about the granitic gneiss. This might be reconciled by assuming (following Dalziel), that the Meall Mor fold and Cona Glen fold (MMF and CGF) are lost in the high strain zone in the western part of the granitic gneiss. The Z-profile folds in the east would verge towards the Beinn an Tuim synform (BTS). Alternatively, it could be argued that the zone of low $D2_g$ strain in the granitic gneiss represents a pair of complementary fold cores, rather than a single core; this would involve disregarding the S-profile folds of lits in the west. This might be favoured by the observation that the lits were not parallel to banding prior to $D2_g$; their orientation is such that S-profile ptygmatic folds would be expected to form on shortening normal to banding (cf. Plate 12.1(c)), and in fact, some probable examples were ween west of (d), in the zone of $F2_g$ Z-folds.

The presence of shear zones at (d) is noteworthy, since if $S2_g$ was tilted on a NNE axis, to make $F2_g$ folds upright and steeply NNE-plunging, the shear zones would become NW-SE, sinistral, and NE-SW, dextral - a situation identical to that at Loch Quoich (section 11.2.1).

The $F3_g$ folds show a similar variation to the $F3_q$ folds at Loch Quoich, in that they are open and gently-plunging in the east, and become tight and steeply-plunging in the west; again this transition marks the steep belt/flat belt boundary, and the Loch Eil Division (Glen Garvan Psammite) appears at a higher structural level than the Glenfinnan Division. Pre-existing variations in bedding orientation could cause the steeper $F3_g$ plunges, but it is the tightness of the $F3_g$ folds, as much as their steep plunges, which distinguishes the steep belt. In addition, folds of $S2_g$ which presumably was relatively constant in orientation prior to $F3_g$, also vary in plunge. Figure 12.2 shows a schematic section across the traverse; the main difference from Loch Quoich is that here, the granitic gneiss appears to contain an $F2_g$ fold core.

There is little evidence of the metamorphic history in this area; $S1_g$, the $MP1_g$ pegmatites, $S2_g$ and the regionally migmatised metasediments, are all identical to their counterparts at Loch

Quoich. None of the late remobilisation was seen here, so late Caledonian metamorphism may have been at somewhat lower temperatures. Again, structural evidence suggests that the granitic gneiss was emplaced as a pre- to early-tectonic intrusion; further arguments will be reserved for Chapter 17.

CHAPTER 13

Strontian

13.1 Introduction

The Strontian intrusion is a forceful Newer Granite (Read 1961), is zoned from a tonalitic margin to an adamellite core, and lies close to the Glenfinnan Division/Loch Eil Division boundary in western Ardgour (Fig. 2.1; Munro 1965, 1973). It postdates the microdiorite suite (Smith 1979; cf. sections 11.1, 12.1), has been dated by U-Pb zircon methods at 435±10Ma (Pidgeon & Aftalion 1978), and is a typical I-type granite (Pankhurst 1979). The absence of hornfels textures in the aureole has been taken to indicate relatively high regional temperatures during intrusion (Watson 1964).

The country rocks, typical striped pelites, semipelites and psammites of the Glenfinnan Division, are coarse-grained and generally migmatitic. Cordierite is present in the inner 300m of the aureole. Ashworth & Chinner (1978) have suggested that this was in equilibrium with garnet, and that the migmatitic lits developed by partial melting. A brief investigation was made of the western aureole at Sron na Saobhaidh [777606], to test this hypothesis.

13.2 Structure

Regional structures appear broadly similar to those encountered elsewhere in the Glenfinnan Division. Bedding and migmatitic banding are deformed in tight-to-isoclinal folds with an axial planar foliation, and this is folded by variably plunging, upright structures, which carry a crenulation cleavage (cf. Brown et al. 1970, and interference petterns on I.G.S. 1-inch sheet 52 in the vicinity of Beinn Resipol [766655]; also cf. $F2_g$, $F3_g$ at Glenfinnan, section 12.3).

Structures in the granite aureole are best seen in coastal exposures, but fresh specimens were collected from the new road cuttings. About 300m west of the granite, tight folds, which verge towards an antiform to the east, and have highly strained long limbs, become common (Plate 13.1(a)). These folds tighten towards the granite, become reclined, and develop an axial planar fabric. In the inner 200m of the aureole, semipelites carry a strong, vertical, planar fabric, which is parallel to the granite margin, and wraps dismembered quartzite bands and basic dykes (Plate 13.1 (e), (f)). There is a sharp contact with strongly foliated tonalite, which contains augened xenoliths, plagioclase phenocrysts and boudinaged basic dykes. Further into the granite, the foliation is weaker, and the dykes are defined by trails of blocky xenoliths veined by granitic material. Thus the inner parts of the intrusion were at least partially molten while the aureole was being deformed.

13.3 Metamorphism

Outside the aureole, the Glenfinnan Division rocks show quartz-plagioclase-biotite-garnet-muscovite-ilmenite mineral assemblages. No significant difference from other areas (e.g. Acharacle) was observed. Dalziel & Brown (1965) record the presence of regional sillimanite, while Stoker (1980) has described regional K-feldspar+sillimanite assemblages from Glen Tarbert, to the east of the granite. The aureole rocks must have suffered an early (?Precambrian) sillimanite-grade metamorphism and migmatisation, comparable to that of the rest of the southern Glenfinnan Division, although possibly at slightly lower pressure or higher temperature than was generally attained.

Within the aureole, cordierite and sillimanite become common, at the expense of garnet and muscovite. Magnetite appears to have replaced ilmenite, and in the innermost aureole, some muscovite-free K-feldspar+cordierite+sillimanite assemblages occur. There is no evidence that these rocks have unusual compositions (section 16.1), so it must be assumed that garnet and muscovite were consumed in cordierite+sillimanite-producing reactions. This is confirmed by the presence of relic garnet in cordierite pseudomorphs (Plate 13.1(b), (c)) - this garnet is very Mn-rich, and may have been stable during thermal metamorphism (section 15.2.3(viii)).

These conclusions are in broad agreement with those of Ashworth & Chinner (1978), as far as mineral parageneses are concerned. These authors also suggested that partial melt migmatites were produced in the aureole. Even at the high temperatures estimated by Ashworth & Chinner (c. 700°C), conditions would be marginal for melting of the trondjemitic lits, especially with $P_{H_2O} < P_{load}$ - cf. Fig. 15.16. In the outer aureole, lits are clearly folded, but closer to the granite, folds have been eliminated by the high ductile strain. Feldspar in the lits is often euhedral, but quartz is deformed (Plate 13.1(d)), and the abundant feldspar porphyroblasts in the inner aureole give clear evidence for subsolidus recrystallisation of this phase (Plate 13.1(g)). Since the larger porphroblasts are always more strongly augened than the smaller, it may be that the smaller ones are later - i.e. that feldspar continued to grow during deformation of the aureole. The semipelites must have been solid (rather than a mesoscopic mixture of melt and restite) during the deformation which boudinaged the quartzites and amphibolites in Plates 13.1(e) and 13.1(f), or the finely banded lits and selvedges could not have maintained their integrity. Tyler & Ashworth (1982) accept that the majority of the migmatitic lits in the aureole represent deformed regional migmatites, but suggest that those K-feldspar-bearing semipelites which have suffered severe ductile deformation underwent minor partial melting, with an intergranular melt film facilitating plastic deformation (cf. van der Molen & Paterson 1979; shear zones in the granitic gneiss, section 17.3.2). Any textural evidence for such a melt has been eliminated by recrystallisation.

13.4 Conclusions

The Late Caledonian Strontian granite was forcibly emplaced into high grade migmatitic gneisses of the Glenfinnan Division. The central parts of the intrusion were at least partially molten during this period, although the margins had probably wholly or very largely solidified. In the aureole, extremely high ductile strains resulted from the forceful emplacement, and cordierite and sillimanite grew at the expense of garnet and muscovite. However, the latter two minerals may still have formed part of the stable assemblage. Although it cannot be completely ruled out, there is no evidence that thermal metamorphism caused partial melting in the majority of the metasediments, certainly not in the typical migmatitic pelites, and the simplest explanation for their migmatitic texture is that pre-existing gneisses were deformed and metamorphosed, with subsolidus development of new mineral assemblages.

CHAPTER 14

Summary of field relationships

14.1 Introduction

Eleven areas have been described (Fig. 14.1). Mallaig and Inverie represent the western Morar Division, with a major slide being present in the latter area; Acharacle and Kinlochourn straddle the Sgurr Beag Slide, with emphasis on the Glenfinnan Division, while Ben Klibreck lies on its probable northward continuation. On Mull and Sguman Coinntich, later folding results in Glenfinnan Division rocks reappearing west of their main outcrop, while Lochan Coire Shubh lies well within the Glenfinnan Division. The Loch Quoich and Glenfinnan sections lie on the Loch Quoich Line, and include outcrops of the West Highland granitic gneiss.

The Morar Division is always of lower metamorphic grade than the adjacent Glenfinnan Division, with a break in grade occuring at the Sgurr Beag Slide. Grade in the Morar Division rises from garnet in the west to kyanite in the east, while that in the Glenfinnan Division is fairly uniformly kyanite/sillimanite. The Loch Eil Division follows on stratigraphically from the Glenfinnan Division with no structural or metamorphic break; there may be a slight upwards decrease in grade within the Glenfinnan Division (cf. P, T estimates in Chapter 15).

In the westernmost Morar Division (Mallaig and Inverie) the peak of garnet-grade metamorphism was reached between MP1 and MS2 and S2 is the dominant fabric, but matrix minerals and microcline porphyroblasts recrystallised during MP2. These textures were overprinted by bedding-parallel high-strain zones, with recrystallisation to biotite-grade (D3) or chlorite-grade (D4) assemblages. The Knoydart Slide is marked by a broad D3 zone, and places kyanite-grade migmatites over the upward-younging, garnet-grade Morar Division. The high-strain zones are folded with bedding and S2 in the core of the upright N-S Morar Antiform, suggesting that at least the D3 zones were originally thrusts (the later ones could be steep shear zones).

In the eastern Morar Division at Acharacle and Kinlochourn, metamorphic grade is higher than at Mallaig (based on geothermometry), and garnets postdate at least one fabric, and are augened by a lower-grade fabric associated with the Sgurr Beag Slide. The presence or absence of two pre-slide fabrics cannot be confirmed due to the higher strain and metamorphic grade. At Loch Eilt (Fig.14.1), suitable Morar Division lithologies are migmatised, and the lits are deformed by fabrics related to the Sgurr Beag Slide.

In the Knoydart Pelite, there was some MP1 garnet growth, followed by a strong S2 fabric and MP2 garnet growth and migmatisation. These textures were deformed by the D3 Knoydart Slide fabric.

In the Glenfinnan Division of Acharacle and Kinlochourn, garnets, migmatitic lits (of early migmatites) and a tectonic fabric are deformed by the lower-grade slide fabric; this is folded by major upright structures with curvilinear axes. Some remobilisation and pegmatite intrusion postdated sliding and continued through the later folding ("late migmatisation"). The Glenfinnan Division regionally overlies the Morar Division, although this is locally complicated by the curvilinearity of the later fold axes.

On the Ross of Mull to the west, a Glenfinnan Division outlier occupies the core of a synform; two sets of early folds are deformed with MP2 migmatites and garnets by a biotite-grade D3 slide, which is folded by the upright NNE Assapol Synform. The Morar Division is deformed by and youngs into the slide.

At Sguman Coinntich to the north, textural features are similar (kyanite-grade migmatites separated from the Morar Division by a later slide, which is itself folded), but the Glenfinnan Division appears to occupy the core of a reclined SE-plunging antiformal sheath fold. this suggests complex refolding, possibly involving two sets of post-slide isoclinal folds.

On Ben Klibreck, there are several slides and Lewisian inliers, but sillimanite-grade migmatites are thrust over Morar Division psammites, with the slide fabric augening lits and garnets in both divisions. The slides are folded by reclined ESE-plunging structures associated with a strong LS shape fabric.

In the Glenfinnan Division of Lochan Coire Shubh, early migmatites (local MP1) are deformed by a strong fabric and isoclinal folds, which are refolded by upright, curvilinear N-S structures.

Further east at Loch Quoich and Glenfinnan, early migmatites and the granitic gneiss have MP1 pegmatites deformed by isoclinal folds with a strong fabric, and refolded by upright, curvilinear N-S structures. These later folds define the Loch Quoich Line (by being tight in the Glenfinnan Division, resulting in steep bedding dips, and open in the Loch Eil Division, resulting in gentle dips). The Loch Eil Division and the granitic gneiss carry all the structural and metamorphic features of the Glenfinnan Division, and there is no high-strain zone like the Sgurr Beag Slide at this boundary.

At Strontian, early migmatites deformed as at Acharacle are further deformed in the aureole of the forcefully-intruded granite.

The Morar Division therefore underwent a low-grade, low-strain D1 event, garnet-grade MP1 metamorphism, a moderately high-strain garnet-grade D2 event and MP2 recrystallisation. The metamorphic peak may have been later (MP2) in the higher-grade areas to the east. The Glenfinnan Division had its metamorphic peak and migmatisation pre-sliding, and folding and renewed metamorphism post-sliding. Sliding was at least D2 in the Glenfinnan Division sequence, but in the areas of least post-slide folding and metamorphism, migmatisation was MP2 and sliding D3, with no significant post-migmatisation, pre-sliding event. The slides were probably originally thrusts (section 14.2), and have been refolded by upright folds in the south and overturned folds (related to D2 in the Moine Thrust Zone) in the north.

Thus both the Morar Division and the Glenfinnan/Loch Eil divisions had an extensive history of deformation and metamorphism prior to their juxtaposition by the Sgurr Beag Slide and subsequent folding and metamorphism. There is no reason to doubt that they formed part of the same pre-slide metamorphic complex, and this is supported by the detailed textural similarities between the western Glenfinnan Division and Knoydart Pelite, and the Knoydart pelite and the Morar Division.

Structural correlations between areas, using slide-reworking of high-grade metamorphic rocks or migmatites as a marker, will be given in section 14.3; firstly, the nature of the slide itself (which bears on its viability as a marker) will be discussed.

14.2 The Sgurr Beag Slide

14.2.1 Introduction

The Sgurr Beag Slide was defined by Tanner (1971) at Kinlochourn, and extended by Tanner et al. (1970) from southern Ross-shire to north of Loch Eilt on the basis of regional mapping. It separates the striped and pelitic, highly-migmatitic Glenfinnan Division from the dominantly psammitic, only locally-migmatitic Morar Division. Lithological criteria for distinguishing the divisions are well-established (section 2.1; Johnstone 1975), and structural criteria for recognising the slide have been described by Rathbone (1980) and Rathbone & Harris (1979). These involve an increase in strain in psammites underlying the slide, attested to by elimination of cross-beds, rotation of folds into parallelism with the extension direction and their subsequent elimination, and finally production of platy psammites in which banding, veins and tectonic fabrics are all parallel.

Rathbone & Harris (1979) suggested that the Sgurr Beag Slide extended as far south as Salen in Ardnamurchan. In the present study, outcrops of the slide were visited at Kinlochourn and Acharacle/Salen, and at its probable northern continuation on Ben Klibreck. To the west, on the Ross of Mull, the slide has been brought down in a later synform. The Knoydart Pelite is separated from the Morar Division by the Knoydart Slide, which could either be a lower slide, or the refolded Sgurr Beag Slide. A slide surrounds the Glenfinnan Division of Sguman Coinntich, but since it appears to lie in an antiform, its structural position is anomalous.

A structure analogous to the Sgurr Beag Slide has been found at every Morar/Glenfinnan division boundary visited, separating rocks of contrasting lithology and metamorphic grade. It is reasonable to consider that the slide forms this boundary throughout Scotland, and that the Loch Coire migmatites in the north are equivalent to the Glenfinnan Division; the slide is best interpreted as a major thrust separating Morar and Glenfinnan nappes (or Glenelg and Ross-shire nappes - Johnstone 1975).

14.2.2 Observations

Areas in SW Inverness-shire will be described from NW to SE in order of increasing (admittedly post-slide) Caledonian metamorphic grade. Assuming the slide to be a major Caledonian thrust, then this sequence is from shallower, more external levels to deeper, more internal levels. The assignation of the northern localities (Sguman Coinntich, Ben Klibreck) to deeper levels is justified by their higher metamorphic grade, and by the fact that only a relatively thin, stratigr ' low Morar Division succession is present. Lewisian inliers become more common in the north, associated with the Sgurr Beag and lower slides, and the Strathy Complex in the northernmost Glenfinnan Division consists either of autochthonous Lewisian (Harrison & Moorhouse 1976, Rathbone & Harris 1979) or deep-seated granulite-facies Moines (J. R. Mendum, verb. comm.). Fig. 14.2 summarises the observations diagrammatically.

Ross of Mull

Low-grade, trough cross-bedded pebbly psammites (probably Upper Morar Psammite) young towards the slide, which separates them from Glenfinnan Division quartzites, migmatitic pelites and kyanite schists. Folds are absent in the psammites, and deformation is by alternating high-strain and low-strain zones, sub-parallel to bedding. High strain zones lacking cross-beds and a few metres wide occur 400m or more from the slide. The last undeformed psammites are 200m from the slide, and throughout the next 100m, low-strain zones become progressively more deformed until all sedimentary structures are lost, while high-strain zones become progressively more platy (banded on a few millimetres scale, all sedimentary structures and planar discordances lost). From 100-50m, the rocks have variable platiness, and in the last 50m they are uniformly platy.

There is a sharp change (defined within 2-3m) from platy arkosic psammite to platy quartzite of typical Glenfinnan Division type (with beds < 1cm thick). Over the next 50m, quartzite beds increase to 10-20cm, and a few platy pelitic bands appear. At 50m, a thick pelitic gneiss appears. It is platy for 10-20m, with augened lits, but soon becomes virtually undeformed.

Bed attenuation in the Morar Division is 2 - 3 times at 150m, 10 times at 100m, 40 times at 50m, and about 50 times in the inner platy zone; displacement on the half shear zone may be 20-40km (section 7.4.2). The deformation was a plane strain, with vertical, E-W xz-plane. Metamorphic grade was rather low - quartz grain-size is severely reduced, with little recovery, but biotite was still stable; garnet is replaced by biotite, quartz and plagioclase. The slide is folded by a major synform which is associated with little penetrative deformation.

Knoydart (Inverie)

Kyanite-grade migmatitic pelites overlie garnet-grade rocks of the Lower Morar Psammite and Lower Striped Schist on a slide which dips SE at 40°. Thick cross-bedded psammites are preserved 200 m below the slide, but thin pelites carry a strong S3 cleavage, slightly steeper than bedding. Intersection lineations pitch at up to 45° to the NE or SW. From 150-50m, the psammites are deformed, cross-bedding is eliminated and an LS shape fabric is developed, with L pitching SW at 70°. In the final 50m, S3 is near-penetrative, with rare lenses of S2 preserved and wrapped by S3 grain-size reduction zones in which new biotite crystallised.

Above the slide, the migmatitic pelites are platy for 50m and contain augened lits and garnets. From 50-100m, the migmatites are deformed but not platy, and lits carry an LS shape fabric parallel to S3; reclined isoclinal folds are present. Above 100m, S3 is steeper than bedding; thin lits are transposed into S3, but thicker ones are folded. Above 150m, F3 folds are more open and overturned to the NW, with curvilinear axes. F3 appears to be a regional deformation event in the migmatites, and extends well east of the slide.

Acharacle (Salen)

The slide is near-vertical, and strikes N-S on the limb of a later upright $F3_a$ fold; it separates migmatitic psammites and pelites of the Glenfinnan Division from non-migmatitic Morar Division psammites. The latter are platy for several hundred metres to the west, due to superimposed $D3_a$ strain. In the last 50m, these psammites are very platy, with quartz ribbons having $\underline{x}:\underline{y} > 20:1$. Fabrics in the psammites are recrystallised and polygonal, but overprinted by grain-size reduction zones of probable $D3_a$ (post-slide) age.

Migmatitic pelites are platy for 100m east of the slide, with quartz in lits recrystallised to 0.1mm grain-size, and feldspars with $\underline{x}:\underline{y} > 10:1$ being common. $F2_a$ reclined isoclinal folds appear 100m into the Glenfinnan Division. Some 400m into the Glenfinnan Division, the boundary between a pelite which occupies the core of a major $F2_a$ isocline, and the underlying migmatitic psammite, is marked by a 50m-wide $F2_a$ platy zone. Subgrains appear in the quartz matrix, and quartz in lits is reduced from several centimetres to less than 1mm. Minor $F2_a$ isoclines appear within 20m of the boundary, and are common throughout the pelite.

All features are variably overprinted by $F3_a$ folds and associated grain-size reduction zones. The tightening of $F3_a$ folds implies that the $D2_a$ slide has been reactivated as a $D3_a$ high-strain zone (as at Garve in the Northern Highlands - Rathbone 1980). In the east of the area, an $MP3_a$ thermal overprint obliterates deformation features while $MS3_a$ -$MP3_a$ pegmatites and local migmatisation become common.

At Loch Eilt to the NE, higher-grade Morar Division rocks are brought up in the core of a major antiform. Both Morar and Glenfinnan divisions are migmatitic, and lits in both divisions are highly deformed for a hundred metres or so either side of the slide (Plate 14.1.(b)).

Kinlochourn

The slide again trends N-S on the steep limb of a later fold and separates non-migmatitic Morar Division psammites from migmatitic Glenfinnan Division psammites and pelites. In the Morar Division, platy psammites with quartz ribbons are present for 150m west of the slide, and the first cross-bedding appears 300m to the west. Fabrics in semipelites are penetrative until 150m, where an earlier fabric is preserved, giving rise to a steep intersection lineation. Slide-age folds ($F2_k$) may be present in the west, but these could be later ($F3_k$).

A strong fabric in the Glenfinnan Divsion augens feldspars and migmatitic lits; reclined isoclinal $F2_k$ folds, possibly with syntectonic garnets, appear 50m into the Glenfinnan Division. The $S2_k$ fabric dies out upwards into a migmatitic pelite - feldspars are augened and wrapped by quartz ribbons in the lower 100m, after which $S2_k$ becomes a crenulation cleavage which augens garnets and individual lits, although internal deformation within the latter is minor. In the higher Glenfinnan Division, $D2_k$ deformation is generally weak, although 10-20m thick platy zones occur.

At Lochan Coire Shubh, a few kilometres to the SE, a map-scale Type 2 interference pattern implies the existence of a major recumbent fold with E-W trending axis, which is folded by upright N-S curvilinear folds similar to $F3_k$, and which on the basis of small-scale features postdates migmatisation.

This fold is probably related to the slide, its E-W axial trend resulting from rotation towards the slide's extension direction, which prior to $F3_k$ would have been E-W or NW-SE. This pattern of originally-recumbent, post-migmatisation F2 folds being refolded by N-S upright structures persists as far east as Loch Quoich.

The slide itself is refolded by upright, isoclinal $F3_k$ folds, some of which show sheath-forms with near-vertical extension directions. Microscopic textures are thoroughly recrystallised and $MS3_k$-$MP3_k$ pegmatites extend across the slide, but no syn-slide pegmatites were recorded. Caledonian metamorphic grade was higher than at Acharacle (perhaps 555°C at 4.1kb vs. 529°C at 3.2kb - section 15.2.3).

Sguman Coinntich

Migmatitic pelites occupy the core of a reclined $F3_s$ sheath fold with axial plane dipping SE at about 40°, and lie 6km to the west of the main Glenfinnan Division outcrop (cf. Figs. 8.1, 14.1), but syn-sliding metamorphic grade is similar to or higher than that at Kinlochourn. Note that if the identification of the $F3_s$ fold as an antiform is correct, then Sguman Coinntich represents an originally deeper level than Kinlochourn (Fig. 14.4). This typical Glenfinnan Division pelite is surrounded by an $F2_s$ high-strain zone similar to the Sgurr Beag Slide.

The Morar Division throughout this area is quite highly deformed, particularly on $F3_s$ fold limbs. However, in the major $F3_s$ fold core, Morar Division psammites within 50m of the $F2_s$ slide are platy, after which rare reclined isoclinal $F2_s$ folds appear. The rocks are flaggy with rare folds for the next 250m, beyond which flaggy and open-folded zones alternate, the latter preserving deformed cross-beds.

The slide is marked by a sharp boundary between psammite and migmatitic pelite, although lenses of both lithologies occur in a 2m-wide mixed zone (Plate 14.1(a)). The migmatitic pelite carries a platy $S2_s$ fabric near the slide, with lits and garnets being augened, but about 50m into the pelite this gives way to a crenulation cleavage which is axial planar to reclined isoclinal folds of bedding and lits. Minor structures are generally recrystallised.

Within the Morar Division, several shells of psammite have been defined by the I.G.S., their boundaries marked by trails of small Lewisian inliers. No marked increases in strain or changes in metamorphic grade were noted in the vicinity of these boundaries, although where they are exposed on the eastern limb of the Sguman Coinntich fold, this may be obscured by the general high strain and later metamorphism and migmatisation. They are pobably lesser slides (cf. those below the Sgurr Beag Slide on Ben Klibreck), or ones in which brittle mechanisms dominated.

Local $MS3_s$-$MP3_s$ migmatisation is common around the eastern margin of the pelite, and intensifies to the east, but $MS2_s$ migmatisation (local segregation) is common in and around the slide zone.

Ben Klibreck

There are numerous slides in the Morar Division, some marked by the presence of Lewisian inliers. Psammites exhibit a 50m-thick platy zone containing rare isoclinal folds, but the platiness dies out 10m into the Lewisian. The latter carries a strong LS shape fabric in acid gneisses, quartz ribbons several centimetres long wrapping recrystallised, elongated feldspars. The platy zones can be traced in the psammites, even where Lewisian is absent, so are not due simply to the lithological contrast. This fabric is folded by $F2_b$ = regional Sutherland F2 (Chapter 9 - reclined isoclinal folds with 30/120-dipping axial planes); the slides are probably also folded on a map scale.

The main slide has a very thick platy zone (several hundred metres) and emplaces migmatitic Glenfinnan Division above Morar Division. It is somewhat obscured by $D2_b$ strain and widespread late ($MP2_b$) pegmatite emplacement. $F2_b$ folds tighten towards the slide, which was reactivated as a $D2_b$ high-strain zone. A Lewisian inlier in the Glenfinnan Division may rest on a higher slide, or could occupy a modified fold core. Microstructures are largely recrystallised by the $MP2_b$ metamorphism, although grain-size reduction from several millimetres to 0.2mm is preserved in the Lewisian.

14.2.3 Conclusions

The varying pattern of structures associated with the slide is shown in Fig. 14.2. The common features in all these areas are that the slide separates high-grade rocks of Glenfinnan Division lithology from lower-grade rocks of Morar Division lothology, and that there is always a sharp break between the two. On the ground, it is always parallel to bedding, but regionally stratigraphic units are cut out - e.g. 200m of Glenfinnan Division pelite is eliminated over several kilometres at Acharacle (Fig. 6.1), and from Lochailort to Kinlochourn the Upper Morar Psammite and most of the Morar Schist have been cut out. A thick zone of high strain is invariably present in the Morar Division, culminating in about 50m of platy rocks adjacent to the slide. The Glenfinnan Division also shows a strain gradient towards the slide, but of variable thickness, it being thinnest where pelites (which would readily deform by slip along the foliation) are present at the slide. These two gradients are represented in Plates 14.1(c) - (e); 14.2(a) - (e); 14.3(a) - (e).

Except in the most westerly area (Mull), slide-age folds (indicating more penetrative ductile deformation) are widespread in the Glenfinnan Division, although only from Acharacle east were large-scale, originally-recumbent isoclines identified. However, even these are small in comparison with the slide, and of lesser extent - i.e. the slide is the primary structure, not the folds. At the deepest levels, there may be some approach to fold-nappe character if the Lewisian lying a short distance above the slide at Kintail (Clifford 1958a) and Ben Klibreck (Chapter 9) can be regarded as occupying an isoclinal fold core.

Deformation in the Morar Division is usually by high-strain zones (bedding-parallel shear zones or minor thrusts). In the deepest areas (Sguman Coinntich, Ben Klibreck) discrete lower slides defined by the presence of Lewisian inliers seem to replace these more-widespread moderately-high-strain zones.

Minor folds were only found in the deepest Morar Division, and no major slide-age isoclines were identified (regionally, reversals in facing or younging in the Morar Division only occur across the Precambrian recumbent folds in SW Inverness-shire - e.g. Powell 1974).

Syn-sliding metamorphic grade increased to the east - from biotite-grade in Mull and Knoydart, through biotite or garnet-grade at Acharacle, to garnet-grade at Kinlochourn, with local migmatisation first appearing at Sguman Coinntich. This gradient is reflected in the fact that, on the whole, deformation in the migmatites at Inverie and Acharacle was localised in micaceous layers, leaving quartzofeldspathic augen preserved, while at Kinlochourn, quartz in lits was deformed into long ribbons which wrap feldspars, and on Ben Klibreck, both quartz and feldspars were severely deformed - i.e. the ductility contrast between micas, quartz and feldspars decreased to the east.

In general, the metamorphic peak postdated sliding and the subsequent folding, hence the "isograd" of $MP3_a$ migmatisation and annealing which cuts across the folded slide at Acharacle, the post-slide migmatisation at Kinlochourn, and the $MP2_b$ migmatisation and metamorphism on Ben Klibreck. On Mull, metamorphic grade was probably constant through the sliding and later folding; at Sguman Coinntich, grade fell after sliding, the rocks being brought up to a shallower level by the $F3_s$ antiform.

The slide is folded on a large scale by tight-to-isoclinal structures (even the Assapol synform on Mull being essentially isoclinal). However, the degree of shortening associated with these folds and the intensity of their minor structures increases to the east. From Kinlochourn southward, they are upright and trend N-S or NNE-SSW, while at Sguman Coinntich and Ben Klibreck their axial planes are moderately SE-dipping. From Acharacle eastwards their axes are highly curvilinear, and an LS fabric with down-dip extension lineation is common. This reflects both higher metamorphic grade and a greater degree of ductile shortening, further into the orogen.

Thus the Sgurr Beag Slide is a major structure which extends sub-parallel to stratigraphy throughout the Northern Highlands, and places high-grade rocks over lower-grade rocks which, however, crystallised at much the same pressure (Chapter 15). It is associated with high LS ductile strains, probably representing at least 40-80km of displacement by simple shear, and is repeated by major upright folds which bring down outliers up to 30km west of the main outcrop. Its initial orientation must have been flat or gently-dipping with E-W to NW-SE extension lineations. Folds are either reclined or overturned to the west, fabrics are parallel to or steeper than bedding and the highest levels of the Morar Division underlie it in the west. The only reasonable interpretation is that it is a major thrust with westerly or northwesterly translation of the hanging-wall. Since the easternmost Morar Division is quite distinct from the westernmost Glenfinnan Division, the displacement must be greater than the cross-strike distance between the most easterly and most westerly outcrops of the slide - at least 50km in Ross-shire and SW Inverness-shire, even without allowing for shortening by later folds.

However, there are differences from a typical high-level thrust zone such as the Moine Thrust Zone. Displacement seems largely to be localised at one horizon, rather than dispersed along several minor thrusts or imbricates, although the presence of numerous small high-strain zones (rather than a few distinct slides) below the Sgurr Beag Slide on Mull and in the Morar Division at Mallaig along strike may represent an approach to this shallower character. Much of the displacement was by ductile deformation (simple shear?) of the wall rocks, especially the Morar Division psammites, while the high-strain zone is much more thoroughly recrystallised than most mylonite zones, and probably formed at higher temperatures. Finite strain may be lower, although the data of section 7.4 could be interpreted in terms of shear strains of several hundred. Recumbent folds related to the slide may have developed at deeper levels in the Glenfinnan Division - these are not found above the Moine Thrust (McClay & Coward 1981). They are known from other high-level nappes

- e.g. the Helvetic Nappes (Ramsay 1981) or the Tay Nappe (Johnson *et al.* 1979), but in both these cases recently-deposited sediments are involved rather than a pre-existing metamorphic complex. The thermal history is different - in general, the Sgurr Beag Slide formed in a regime of increasing temperatures, and conditions became more ductile as deformation proceeded, rather than less ductile as in the typical thrust sequence from mylonites to brittle faults (e.g. Butler 1982a). In fact, from the arguments of section 2.2.1(iv), brittle faulting probably preceded the main ductile movement in the slide zone. The ubiquitous folding of the slide reflects this regime: as temperatures rose, the strain-softened rocks in the slide zone would be recrystallised while the country rocks became weaker, and deformation could then become more general. By the time temperatures fell low enough for brittle faulting, the slide was thoroughly recrystallised and refolded, and behaved mechanically as a stratigraphic boundary.

Within the Glenfinnan Division, the down-structure increase in penetrative deformation associated with the slide is probably due to increasing temperature and overburden. A similar gradient occurs in the Morar Division, but deformation is always much less penetrative. This could result from the lithological contrast (massive psammite *vs.* striped pelite/semipelite/psammite) but there may also be some thermal effect. Although the Morar and Glenfinnan Divisions had essentially cooled from their Grenville metamorphic peaks by the time sliding began (cf. the preservation of Precambrian radiometric dates), the Glenfinnan Division may still have been slightly hotter, resulting in a locally-inverted thermal gradient (e.g. Powell *et al.* 1981). Also, the Glenfinnan division rocks which overlie the presently-exposed Morar Division must have suffered the early increments of their slide deformation at greater depth, or at least further east where the Caledonian geothermal gradient was probably higher. It might be expected then that the folds in the Glenfinnan Division formed early in the transport of the Glenfinnan/Loch Eil nappe, and were modified by the high strains in the slide zone as the nappe climbed to higher levels - hence the restriction of platiness to only one limb of the $F2_a$ isocline lying above the slide at Acharacle. However, before the rocks presently exposed reached a level at which cataclasis could occur, the Sgurr Beag Slide ceased to move and was overtaken by the main Caledonian thermal event. Figure 14.3 shows the probable distribution of slide-related structures prior to refolding.

Whether the unmodified nappe depicted ever existed is relevant to the use of the slide as a time marker. If movement on the Sgurr Beag Slide ceased on initiation of a deeper thrust, with the slide folded in response to movement on this new thrust, then two deformation events are represented in the classical sense (Fig. 14.4). If the slide locked at its deeper end for metamorphic reasons, or due to increasing overburden, then both sliding and refolding would be diachronous, but would still form a coherent, correlatable sequence for descriptive purposes.

The first model is favoured by the fact that the structures which fold the slide in Sutherland are typical F2 folds, correlated by Soper & Brown (1971) and McClay & Coward (1981) with mylonitisation in the Moine Thrust Zone - i.e. when movement on the Moine Thrust began, that on the Sgurr Beag Slide ceased, and the slide was folded along with the rest of the Moine Nappe. The presence of lesser slides within the Morar Division may mean that this distinction is less abrupt in detail, but at least those nearest to the Sgurr Beag Slide could be related to duplexes, and so broadly contemporaneous with it (cf. the Knoydart Slide discussed below).

If slides further west are of D2 age (as suggested by Mendum 1979), and only those close to the Sgurr Beag Slide are $D1_k$, then the Sgurr Beag Slide and its related structures must have formed the Caledonian deformation front during $D1_k$, with the Morar Division being virtually undeformed (except for a few minor bedding-parallel shear zones as at Mallaig) until movement was transferred to the Moine Thrust during $D2_k$.

The analogy between the Sgurr Beag Slide and a shallow thrust comprising ramps and flats, which always cuts up-section and places deeper (older) rocks on shallower, and in which more external thrusts formed later (Dahlstrom 1970, Boyer & Elliott 1982) has to be applied with caution. For instance, where the slide transected earlier folds, it could locally cut down-section, and a pre-existing metamorphic complex would not show the same mechanical anisotropy as a layered sequence of sandstones, limestones and shales. The isoclinal sheath-like later folds could not just be produced by ramping of a deeper thrust, and must reflect some regional simple shear or ductile shortening affecting the Moine Nappe as a whole.

As an example of how such a model might be applied, consider the Knoydart Pelite (Chapter 4, Fig. 14.5). The Knoydart Slide shares the structural characteristics of the Sgurr Beag Slide, and fits logically into the sequence in Figs. 14.2 and 14.3. Although metamorphic grade is higher in the Knoydart Pelite than in the adjacent Morar Division, its lithology is still typical of the Morar Division, and its metamorphic grade is lower than that of the nearest Glenfinnan Division, but similar to that of the Morar Division at Kinlochourn, 20km to the east (section 15.2). The Sgurr Beag Slide at Kinlochourn places Glenfinnan Division on the Lower Morar Psammite or the lower part of the Morar Schist. The Knoydart Pelite overlies the same horizon, and has a platy boundary with the attenuated Lower Striped Schist (Tanner 1971, Powell 1974 and I.G.S. 6-inch maps). Across the Morar Antiform at Mallaig, there is a continuous sequence through the Morar Division, while along strike at Lochailort and the Ross of Mull, the Glenfinnan Division lies on the Upper Morar Psammite in the core of a synform (Fib. 14.5(a)).

Suppose that in the vicinity of Kinlochourn there was a ramp on the Sgurr Beag Slide where it climbed from the Lower Morar Psammite to the Upper Morar Psammite (Fig. 14.5). The Knoydart Slide then develops further west, forming a Knoydart Duplex.* On this model, the Knoydart Pelite is the higher-grade lateral equivalent of the Morar Schist, and the Upper Morar Psammite

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* A duplex, in the sense of Butler (1982b) and Boyer & Elliott (1982), is a family of horses (minor thrust slices) bounded by a floor thrust and a roof thrust. Only floor and roof thrusts are shown in these sections, but the scale of rock-units such as the Knoydart Pelite compares more favourably with duplexes than with horses, and there are probably minor slides within the unit which cannot be identified in the absence of good stratigraphic markers.

correlates with the Aonach Sgoilte Psammite of Ramsay & Spring (1962). If these quartzites were considered to be part of the Glenfinnan Division (G.S. Johnstone, verb. comm.) a small earlier duplex could be inserted (position dotted) giving rise to final model II in Fig. 14.5. This has been drawn with the early duplex transported into the thick psammitic sequence at Arisaig; alternatively, it might not be exposed at present.

Clearly this is only a crude model, and it hides its most testable portions above or below ground, but it does show some promise. The Knoydart Slide would be broadly synchronous with the later movement on the main Sgurr Beag Slide, although the branch of the Sgurr Beag Slide forming the top of the duplex had stopped moving. If like those in the Moine Thrust Zone this duplex had limited lateral continuity, its slide-bounded southern margin may help to explain the complex geology around Loch Beoraid; but note that Powell (1974, fig. 1 — Fig. 14.5(a)) implies reversals of younging in this area, which if genuine call for more than a simple thrust model. The suggestion that a ramp lay close to the present outcrop of the Sgurr Beag Slide in western Inverness-shire helps explain its unusually strong cross-cutting relationships in this area - the slide transgresses from the Lower Morar Psammite to the Upper Morar Psammite in a short cross-strike distance, and according to Baird (1982), cuts out a map-scale fold. This is probably not the case further north, where more of a "layer-cake" tectonic stratigraphy appears to be present (cf. the "shells" of psammite around the Sguman Coinntich and Beinn Dronaig pelites, Chapter 8), as would be expected closer to the original décollement. Since deeper levels of the slide are exposed in the north, its original strike must have been more N-S than at present, which together with the E-W trend of early isoclines suggests that the slide was more westerly-directed than the Moine Thrust.

The strain-based criteria suggested by Rathbone (1980) for definition of the slide have been confirmed and extended into the Glenfinnan Division. However, in defining the relative magnitudes of slides, Lewisian inliers are not a reliable indicator. Their presence confirms that large displacements have taken place, but their absence does not imply a smaller displacement. For example, the Lewisian inliers associated with minor slides in the Morar Division of Sguman Coinntich and Ben Klibreck could be autochthonous basement to this stratigraphically low Morar Division, and hence much less far-travelled than the Glenfinnan Division, with or without its own Lewisian basement (Fig. 14.4). No-one would suggest that imbricates in the Moine Thrust Zone which involve Lewisian are further-travelled than the overlying Moine Nappe itself. A metamorphic and stratigraphic contrast between rocks of broadly the same age is the best criterion to use when searching for major nappe boundaries (cf. the facies belts which define the Alpine cover nappes - H. P. Laubscher in McClay 1981). Lewisian inliers can only be introduced by the mechanism of Fig. 14.4 where the slide lies at a level close to the Moine/Lewisian boundary. In W Inverness-shire, where it lies well above the Glenelg and Morar inliers, any duplexes will only involve Moine rocks - this need not imply smaller displacements, particularly where, as in Knoydart, there is a metamorphic contrast across a slide.

An alternative model for the introduction of Lewisian inliers has been proposed by Tanner et al. (1970), Johnstone et al. (1979) and Rathbone (1980). These authors envisaged a basement ridge separating Morar and Glenfinnan division depositional basins, which was intersected by Caledonian thrusts to generate the Lewisian inliers. A simple thrust model cannot readily be used to inject "pips" of Lewisian into slide zones - this would simply be a smaller-scale version of Sutton & Watson's (1962) discredited "cicatrice" model for inserting Lewisian inliers into the Moines as a whole. Lewisian inliers within the Glenfinnan Division are readily explained as fold cores or as basement to the local Glenfinnan Division, resting on subsidiary thrusts. However, a stacking of Morar Division, Lewisian, Morar Division, Lewisian, Morar Division, Glenfinnan Division implies either that each unit represents a separate nappe, and the highest Morar Division slice originally lay east of the supposed basement high, within the Glenfinnan Division basin, or that each Lewisian inlier represents autochthonous basement to the overlying Morar Division.

Whichever model is preferred, shallow basement must have lain west of the Glenfinnan Division prior to sliding. The question is whether this was inherited from the original form of the Moine sedimentary basin, or is the product of Precambrian (Grenville?) tectonism. Fib. 14.6(a) shows the required disposition of major stratigraphic units in the footwall of the Sgurr Beag Slide (the hanging-wall consisting essentially of Glenfinnan Division). On restoring the minor thrusts, Fig. 14.6(b) is produced. Fig. 14.6(c) shows a purely tectonic model for the shallow Lewisian, Fig. 14.6(d) one involving an inherited basement high. The requirement for Morar Division rocks to overlie or lie east of the basement high in Fig. 14.6(d) weakens the arguments for separation of the two basins - one might have expected these Morar Division slices to have some Glenfinnan Division characteristics, and Glenfinnan Division lithologies would have to be restricted to areas well east of the ridge, otherwise they should be included in the slices which underlie the main slide. It is also rather surprising, if the high formed a source for Glenfinnan and Loch Eil division sediments (Johnstone <u>et</u> <u>al</u>. 1979), that it is overlain by Lower Morar Psammite - one might have expected the Morar Division sediments to <u>onlap</u> this high, as in Fig. 14.6(e), with the lower units pinching out well to the west of it. Such an arrangement could not readily produce the present structure. The Lower Morar Psammite might be diachronous, the proximal equivalent of the other units (Fig. 14.6(f)), but it is the Upper Morar Psammite in the extreme west which shows the most convincing shallow-water features. A model such as Fig. 14.6(g) could produce the present structure, but it would require the Glenfinnan and Loch Eil divisions to be either younger or older than the Morar Division (in the former case, the Morar Division was eroded off the developing high prior to or during Glenfinnan Division deposition). Presumably, uplift would have been smaller in the south, allowing sediments to cover the basement high, so that the slide cut through a higher level of the Morar Division statigraphy. The association between shallow-level characteristics of the Sgurr Beag Slide and high stratigraphic levels of the Morar Divison would be pure coincidence, since in Fig. 14.6(g), the same level of the slide intersects different levels of the earlier matamorphic complex.

On the model of Fig. 14.6(c), syn-sliding metamorphic grade, the stratigraphic position of the slide in the Morar Division and the presence of Lewisian inliers are all intimately connected, and reflect the structural level currently exposed. The Glenelg Lewisian could represent minor modifications to a generally flat-lying Moine/ Lewisian interface, in which case the Glenfinnan and Loch Eil divisions could be straightforward lateral equivalents of the Morar Division. There might have been exposed Lewisian sources for some of their sediments, but these need not have had the special location required by the "basement ridge" model. Alternatively, there may have been major Precambrian nappes (dotted in Fig. 14.6(c)), with the Morar and Glenfinnan divisions lying on separate right-way-up limbs. This would allow a large separation between the two (required by their lithological differences and the restriction of early igneous suites to one or other division - Johnstone (1975) and Winchester (1976)), but this can be adequately provided by the tens of kilometres displacement on the Sgurr Beag Slide.

Thus the main conclusion is that the Sgurr Beag Slide is a major westerly-directed thrust, traceable throughout the Moine outcrop and with at least tens of kilometres displacement, which disrupts an older metamorphic compex and places Glenfinnan Division migmatites on lower-grade Morar Division. At the present level of exposure, the slide and its associated slide zone or shear zone is the dominant structure, with contemporaneous folds, although present, being on a much smaller scale. In the vicinity of its main outcrop, the Sgurr Beag Slide forms the boundary between an eastern area in which early Caledonian deformation was widespread with abundant folding and earlier events are largely obscured, and a western area where early Caledonian deformation was restricted to narrow shear zones or absent entirely, and where relatively large outcrops (tens of metres to kilometres wide) preserve undeformed Precambrian structures. In its westernmost outlier, the slide disrupts virtually undeformed

Morar and Glenfinnan division rocks, implying that at shallow levels it behaved as a discrete thrust with little internal deformation occurring in either unit. Later Caledonian deformation and metamorphism affected the whole outcrop width of the Moine Nappe. Thus the Sgurr Beag Slide at depth effectively formed an early Caledonian deformation front, although it is analogous to the Moine Thrust rather than the Sole Thrust in that it lies at the base of the major allochthonous nappe, rather than being the lowest thrust within the zone of imbricates and duplexes. In the terminology of Boyer & Elliott (1982), the Glenfinnan and Loch Eil divisions form a dominant thrust sheet, with the Sgurr Beag Slide having much greater displacement than those which immediately underlie it. On a larger scale, the entire Moine Nappe forms a dominant thrust sheet, with the Moine Thrust having a larger displacement than the underlying thrusts within the foreland sequence.

14.3 Regional structural correlations

Regional correlations will use the reworking by an early Caledonian slide of a Precambrian high-grade metamorphic/migmatitic complex as a marker. From the preceding discussion, and that of section 2.2.1(iii), this is a realistic marker in the usual sense of structural correlation, and since the Sgurr Beag Slide can be traced throughout the Northern Highlands, it provides a link between widely-separated areas where constancy of such features as fold style and orientation, and simple numbered sequences, could not be expected to hold. The resulting correlations are more readily reconciled with published radiometric dates than some previous ones (section 14.4).

Table 14.1 summarises the correlations between field areas visited in the course of this study. There is an apparent decrease in complexity across the table from left to right; late, open structures like those recorded at Malliag are present elsewhere but were not studied in detail, but the decrease in the number of pre-slide phases of deformation does accurately reflect the field observations. In detail, it can be seen that this is a west to east contrast, rather than a Morar/Glenfinnan division one - the Glenfinnan Division on Mull, the Morar Division of Mallaig and Inverie and the Knoydart Pelite all carry two pre-slide fabrics, but in the other areas, only one was recognised in both Morar and Glenfinnan Divisions. The Ben Klibreck sequence could be rewritten to correspond with Sguman Coinntich, since there is good if circumstantial evidence for a pre-slide fabric, but as presented it compares more readily with the standard Sutherland sequences of Soper & Brown (1971) and Soper & Wilkinson (1975). This reduction to one pre-slide fabric can be explained by the intensity of later deformation in these areas: even at Mallaig, where post-D2 strain is very low, D1 is defined mainly by the fact that S2 is a crenulation cleavage. If S2 became penetrative in the higher-grade areas, D1 would be difficult to define anyway, and when slide and post-slide deformation was superimposed, this would be virtually impossible. In the eastern areas, even local D1 (equivalent to the strong S2 of Mallaig) is usually only defined by an early fabric which goes round isoclinal folds. Thus in those areas where two pre-slide fabrice have not been identified, it can reasonably be argued that they would not have been recognised, whether they were present or not. In fact, at each of Acharacle, Kinlochourn and Loch Quoich, one or two interference patterns were seen which suggested the existence of a pre-D1 phase of deformation.

The correlation of the slide has been discussed; the originally-recumbent isoclines in the Glenfinnan Division (Lochan Coire Shubh, Loch Quoich and Glenfinnan) are similar in orientation to those which overlie the slide, and occupy the same position in the structural sequence, so correlation of these with the slide is reasonable. The correlation of post-slide structures is less certain. The F3 folds of Acharacle, Kinlochourn, Lochan Coire Shubh, Loch Quoich and Glenfinnan are all very similar in style and orientation; the similarity between the last two localities is particularly striking, and extends to sets of minor shear zones and conjugate folds in the granitic gneiss. The Caledonian metamorphic peak in the western Glenfinnan Division is broadly syn-D3, while at Loch Quoich it is syn- to post-D2; it may be that more than one set of folds is represented in the classical sense, but they can all be grouped together as post-slide, upright structures. The Morar Antiform and Assapol Synform occupy the same position in this sequence, although they are much more open, and often lack minor structures - in fact, at Mallaig it was impossible to decide which minor structures, if any, were related to the Morar Antiform. In the north, the reclined post-slide folds are associated with a strong LS fabric which at least on Ben Klibreck corresponds to Soper & Wilkinson's (1975) S2, the main fabric in the Moine mylonites. Two sets of tight post-slide folds are required to explain the structure around Sguman Coinntich, possibly corresponding to F2 and F3 of the thrust zone. Precise correlation of these reclined folds with the upright structures further south should perhaps be avoided; although they are both post-sliding, there is no real constraint on their minimum age.

To sum up, the structures in the Moines can be split into a pre-slide set, associated with high-grade metamorphism, the slide and associated recumbent folds, and a post-slide set with renewed metamorphism. On a regional scale, it is probably not worth attempting a more detailed correlation of the members of each set.

Correlation with other areas can best be discussed with reference to Table 14.2, taken from Powell (1974). Shepherd (1973) and Wilson (1975) considered the Sgurr Beag Slide to be part of their D2 event; by analogy with Ben Klibreck (section 9.5.1), the slide probably represents a distinct event, with their D1 corresponding to the pre-slide fabric, and D2 to $D3_s$ and $D2_b$, i.e. to the thrust belt D2. S1/F2 interference patterns in the thrust belt result from progressive deformation in the mylonites (Butler 1982a) and correspond to a single post-Arenig phase in the Moines - F2 of Soper (1971) and Soper & Wilkinson (1975). The latter authors suggested that veins which cut the mylonite banding but carry an L2 lineation prove the separation of D1 and D2, but they could be syn-D2 veins, or the fabric the result of late-D2 homogeneous strain (cf. the Vagastie intrusions, section 9.5.2). McClay & Coward (1981) postulated a distinct mylonite-forming event, but even this was probably restricted to the mylonites, and did not affect the Moine Nappe as a whole. S1 in the Moines above the mylonite zone corresponds to the Precambrian

S1 of Wilson & Shepherd (1979) and not to S1 in the mylonites.

In Glenelg-Arnisdale, F1 and F2 probably correspond to D1 and D2 of Mallaig; the precise assignation of F3 and F4 is more doubtful, but they probably fall into the post-sliding (Morar Antiform and later) group. At Kinlochourn, Tanner's (1971) F2 corresponds to $F3_k$ of this thesis, while the slide, which he did not distinguish as a separate event, is assigned to $F2_k$. The F1 folds from east of the slide which were shown by Tanner (1971) appear to be $F2_k$ or $F2_1$; those west of the slide could be of the same age, or may be equivalent to F2 of Mallaig, i.e. pre-slide. Similarly, Tanner's $F1^a$ represents F1 and/or F2 of Mallaig. There are some disagreements between Tanner (1971) and this thesis (section 5.5) in the assignation of particular folds.

In Moidart-Ardgour, Brown _et al._'s (1970) sequence can be broadly correlated with the Acharacle/Glenfinnan sequences, although there may be detail disagreements in the assignation of specific folds, and particularly over the timing of migmatisation and granitic gneiss formation; Brown _et al._ (1970) and van Breemen _et al._ (1974) erroneously linked these phenomena in Ardgour with the much later emplacement of discrete pegmatites.

Powell's (1974) sequence is difficult to correlate directly, since it attempts to cover a wide area, and some of the correlations within it are considered to be invalid. In the west, his F1 and F2 minor folds and fabrics correspond to D1 and D2 of Mallaig, but major structures (especially the Morar Antiform) are here considered to be post-D2. Powell's western F3 may correspond in part to D3 of Mallaig, but his F3 Glenshian Synform (Fig. 14.5(a)) folds the slide, and so would correspond to $F3_a$, $F3_k$ etc., or to F4 of Mull. It is referred to F4 in Powell _et al._ (1981) and Baird (1982). The shear zones which provided the bulk of D3 (Mallaig) deformation were not recognised by Powell (1974). Powell's (op. cit.) F2 folds in the east correspond to Brown _et al._'s (1970) F2, and so to $F2_a$ etc. Thus the main disagreements with Powell (1974) lie in the age of the major structures in the west (but not of fabrics, garnets etc.) and in the suggestion from the present work that the western D1 and D2 cannot be distinguished in eastern areas, and form a composite D1.

Tobisch _et al._'s (1970) seven-fold sequence is more difficult to fit in - it may be that some of their correlations are wrong, and there should be less phases, or that this approaches the true deformation sequence, and several sets of folds have been wrongly assigned to one phase elsewhere due to their general parallelism and lack of critical exposures in hinge zones: a fold phase that was only observed at one or two outcrops in a small map-area could not reasonably be given a place in the sequence. However, the pre-Cannich folds are associated with the primary migmatisation of the Moines, so might reasonably be correlated with $D1_k$, $D1_s$ etc., and Cannich (long-limbed isoclinal folds with SE-dipping axial planes) with $D2_k$, $D2_a$ and sliding. Their Strathfarrar and Orrin folds occur in separate areas (the former mainly in the Glenfinnan Division, the latter in the Morar Division) and their age relations are not clear. Strathfarrar folds appear to end against the slide, perhaps due to its reactivation as a later high-strain zone - cf. the $F3_a$ folds adjacent to the slide at Acharacle. In style, metamorphism and orientation the Monar folds compare with $F3_s$, $F3_q$ etc., while the later phases correspond to the late structures which were not specifically mapped in the present study. In the east of their area, Tobisch _et al._ (op. cit.) detected a metamorphic low between Cannich and Monar events, and suggested that this represented the hiatus between two orogenies. If genuine, it could be due to a decline in temperatures caused by early Caledonian uplift of the Glenfinnan Division on the Sgurr Beag Slide, followed by prograde metamorphism in the main Caledonian event. Such a temperature decrease might cause increases in ductility contrasts between rock-units, and promote additional folding.

A revised regional correlation is given in Table 14.3. Note particularly how the Carn Chuinneag granite and the 730Ma pegmatites are both placed in the interval between Precambrian and Caledonian deformation events, while the 430-450Ma pegmatites postdate the Sgurr Beag Slide but predate some Caledonian deformation. The c. 1000Ma date from the Ardgour granitic gneiss (Brook et al. 1976) is also consistent if its pre-$D1_g$ age (= F1 of Brown et al. 1970) is accepted, as is the Precambrian date of Brook et al. (1977) for the western Morar Pelite (corresponding to the D2 metamorphic peak of the Mallaig sequence). The implications of published radiometric dates will be discussed more fully below.

The most important feature of this correlation is the assignation of local D2 in the bulk of the Glenfinnan Division to a distinctly later event than D2 in the western Morar Division and Glenfinnan Division outliers such as Mull. It means that (based on the discussion below of radiometric dates) the dominant fabric in the Glenfinnan Division is Caledonian, and that (for example) the Loch Quoich Line is a front not of Caledonian reworking (cf. Roberts & Harris in press) but of late reworking in rocks already thoroughly-deformed during the Caledonian orogeny. This interpretation is implicit in the description of the Sgurr Beag Slide in section 14.2 - if the originally-recumbent F2 folds from Lochan Coire Shubh to Loch Quoich were correlated with F2 at Mallaig, then slide-age deformation would be lacking in the Glenfinnan Division (since the upright N-S F3 folds are not likely correlatives of the slide), which would suggest that it was a discrete thrust within a rigid foreland, perhaps comparable with the Outer Isles Thrust, rather than the base of a major nappe and a front of strong early Caledonian deformation.

In view of the importance of this conclusion, the evidence should perhaps be emphasised. Although the garnet-grade metamorphic peak at Mallaig is MP1, migmatisation in the Glenfinnan Division of Mull and the Knoydart Pelite is MS2-MP2 (although quartz-rich MP1 veins are transposed into S2), and major cross-cutting pegmatites in Knoydart are MP2. It is therefore likely that migmatisation in the Glenfinnan Division was MS2-MP2 in an absolute scale; since it is local MP1 in the main Glenfinnan Division sequence, it is reasonable to suppose that the weak Mallaig S1 cannot be recognised. The c. 750Ma (Morarian) pegmatites dated by van Breemen et al. (1974) from just west of Glenfinnan were MP1 in Brown et al.'s (1970) sequence; this can only be reconciled with the evidence for a Grenville (c. 1000Ma) metamorphic peak if local D2 is Caledonian (or at least post-750Ma) and hence this cannot be correlated with the Precambrian Mallaig D2. The sequence at Loch Quoich is so similar to that at Glenfinnan that it is inconceivable that $D2_q$ is Precambrian, $D2_g$ Caledonian. If D2 (Glenfinnan) and D2 (Mallaig) were correlated, it would force a choice between the reality of Grenville and Morarian events; the correlation of Table 14.1 allows both to be accepted. The reasons for the previous mis-correlation of these two events can be readily understood, since early structural analysis relied on refolding of folds, and platy zones were largely ignored (their significance in the Moines not being fully appreciated until the work of Rathbone 1980). Assuming mapping to start in the lower-grade, apparently-simpler Morar Division, an F4 fold such as the Glenshian Synform would be seen to refold F2 folds; since the D3 platy zones were not recognised, the fold would be assigned to F3, and the refolded slide-age folds in the Glenfinnan Division (actually F3) to F2. Both sets of folds are tight-to-isoclinal with a high-grade crenulation fabric, and would not readily be distinguished on style alone. Baird (1982) and Powell et al. (1981) admit that the Glenshian Synform can only be re-dated as an F4 because it folds the D3 slide, and that some of Powell's (1974) eastern F3 folds should be re-assigned to F4 in an absolute sequence. The correlation suggested in this section would assign most of them to Morar Division F4, and Powell's F2 to Morar Division F3, leaving Powell's F1 to correspond to F1 and F2 of western Morar. Thus the major early Caledonian folds east of the Sgurr Beag Slide correspond west of the slide to discrete narrow shear zones and few if any folds.

This correlation helps to explain one of the more unusual features of the Glenfinnan/Loch Eil division boundary - the so-called "graptolite pelite". Between Loch Quoich and Glen Affric, tongues of Glenfinnan Division pelite extend into the Loch Eil Division and are folded by the F3 structure which defines the Loch Quoich Line, giving rise to hook-shaped interference patterns (Figs. 2.1, 17.1; Roberts & Harris in press, fig. 2). These pelites occupy isoclinal F2 fold cores (A. L. Harris & A. M. Roberts, verb. comm.) On unfolding the F3 structure, F2 folds of the division boundary with wavelengths of several kilometres have a consistent S-profile looking down-plunge to the east. If F2 at Loch Quoich ($F2_q$) was originally recumbent with E-W axes as suggested above, then these major $F2_q$ folds were originally overturned to the north, although they do not appear to verge towards a larger structure. The analogy can be drawn here with later folds developed above the active Moine Thrust Zone (McClay & Coward 1981). Suppose the folds originally formed with their axes perpendicular to the shear direction and were overturned to the west or northwest (cf. F3 in the thrust zone). As deformation continued, they would be rotated towards the shear direction (cf. F2 in the thurst zone). If the fold axes were strictly perpendicular to the shear direction, sheath folds would be produced, alternately overturned to the north and south when viewed down the extension lineation (cf. Sanderson 1973). If the fold axes formed oblique to (clockwise of) the shear direction (e.g. due to pre-existing anisotropies) or if the shear direction changed after fold nucleation (perhaps due to unequal advancement of different parts of the nappe - cf. Coward & Kim 1981) then the final folds would be consistently overturned to the north. It is noteworthy that above the Moine Thrust Zone at Loch Eriboll, the majority of the reclined F2 (thrust zone) folds are overturned to the south (Soper & Wilkinson 1975), perhaps due to the operation of such a mechanism; some are overturned to the north, presumably due to variations in the initial axial trend. Such a mechanism would suggest that the $F2_q$ folds formed in response to simple shear in the nappe overlying an active, westerly-directed thrust - presumably the Sgurr Beag Slide. The alternative correlation (of $F2_q$ with the Precambrian F2 at Mallaig) provides no specific explanation for this systematic northwards vergence. Note that the E - W trend of the $F2_q$ axes does not favour formation in an E - W (Grenville?) orogenic belt - folds as tight as these would certainly have had their axes rotated towards the extension direction, which should trend N - S in that situation.

There are two other important conclusion. One is that the Morar Antiform and associated folds are post-D2 (Mallaig); in this case, the fabrics and minor structures defined by Powell (1974) are accepted, but their relationship to major structure questioned. A corollary of this is that F2 (Mallaig) was originally recumbent (as suggested by Poole & Spring 1974). The fact that recumbent F2 Precambrian structures were reworked in recumbent $F2_k$ etc. Caledonian structures is not a problem, since if deformation in the Glenfinnan nappe involved regional simple shear, even beds lying close to the shear plane could be in the shortening field, and fold axial planes (initially at 45°) would be rotated towards the sub-horizontal shear plane as deformation proceeded. The other significant conclusion is that D2 in the Moine Thrust Zone corresponds to a regional Caledonian event in the Moines, which deforms both the Sgurr Beag Slide (an earlier thrust?) and an earlier Precambrian fabric, and that S1 in the Moine mylonites does not correspond to this Precambrian S1 in the Sutherland Moines.

14.4 Assessment of published radiometric dates

14.4.1 Introduction

Radiometric evidence for a Precambrian event affecting the Moines was first brought forward by Giletti *et al.* (1961) and confirmed by Long & Lambert (1963). These authors obtained Rb/Sr muscovite dates of 665-740Ma (interpreted as minimum ages) from large pegmatites at Knoydart, Sgurr Breac and Carn Gorm (locations of Fig. 14.7). The dates were used by Lambert (1969) to infer the existence of a *c.* 750Ma Morarian metamorphic event. Numerous K/Ar and some Rb/Sr determinations on micas from schists and pegmatites gave dates in the range 400-430Ma and confirmed the importance of Caledonian metamorphism (Giletti et al. 1961, Long & Lambert 1963, Brown et al. 1965, Miller & Brown 1965). In the years since 1961, it had become accepted that the 750Ma dates referred to a real Morarian event; however, the c. 1000Ma Rb/Sr isochron obtained by Brook et al. (1976) reopened the possibility that the younger dates were due to partial Caledonian resetting or slow post-Grenville cooling.

In this section, published radiometric dates will be discussed with respect to the structural correlations outlined previously. The results will essentially be grouped geographically from south to north. A rough recalculation has been made of dates from older papers, using the decay constants of Steiger & Jaeger (1977). This results in a c. 8Ma increase in K/Ar dates in the range 400-450Ma (J. L. Swallow, verb. comm.) and a +3.5% or -2% change in Rb/Sr dates (depending on which constant was used originally). Some of Giletti et al.'s (1961) dates were subjected to analytical corrections by Long & Lambert (1963); these have been further corrected as above. Corrected dates are identified by an asterisk (*).

Caledonian mineral dates are considered on the whole to represent cooling or blocking ages (except perhaps U/Pb zircon) while whole-rock ages date the closure of the system to diffusion on the scale of the sample. Older dates will be partially reset and give minimum estimates for the age of Precambrian events. The blocking temperatures of Purdy & Jaeger (1976) have been used, and are summarised below:

U/Pb monazite	530°C
Rb/Sr muscovite	500°C
K/Ar muscovite	350°C
Rb/Sr biotite	300°C
K/Ar biotite	300°C

At Glen Dessary, van Breemen et al. (1979b) suggested that the blocking temperatures for K/Ar hornblende and U/Pb monazite or sphene were similar, and comparable to that for Rb/Sr muscovite, and that U/Pb apatite compared with K/Ar muscovite. This is an important conclusion, since the coincidence between Rb/Sr and U/Pb dates found by van Breemen *et al.* (1974) was one of the more compelling pieces of evidence for a genuine *c.* 750Ma event.

14.4.2 Summary of available dates

Glenfinnan-Loch Eilt

The most convincing Grenville (*c.* 1000Ma) dates come from the Ardgour granitic gneiss. Brook *et al.* (1976) reported a 1028±43Ma large-sample whole-rock Rb/Sr isochron, with an initial Sr^{87}/Sr^{86} ratio of 0.709 (recalculation by Brewer *et al.* 1979). Pidgeon & Aftalion (1978) found upper and lower intercepts to a U/Pb discordia of 1556Ma and 574Ma; however, Aftalion & van Breemen (1980) reinterpreted both this and new zircon data in terms of initial homogenisation at *c.* 1750Ma, and partial resetting at 1030-1100Ma and *c.* 490Ma. The adjacent Glenfinnan Division metasediments gave similar results, but the Grenville component of disturbance was less important. Thus both these rocks show evidence of strong Grenville and Caledonian events, but no Morarian event.

The granitic gneiss was intruded into the Moines as an igneous body (section 17.3.1) prior to or during $D1_g$ (=D1 of Brown *et al.* 1970 - note the disagreement with their conclusions and those of van Breemen *et al.* 1974, 1978). It carries an $S1_g$ deformation fabric; however, given the probability that $D1_g$ corresponds to more than one deformation event (Tables 14.1, 14.3) it could still be syntectonic with respect to a Grenville orogeny. It could be argued that the granitic gneiss was intruded at 1028Ma and thermally metamorphosed the adjacent sediments, but that $D1_g$ deformation did not take place until later (*c.* 750Ma?); however, it is surprising that no record of this event is preserved in the isotopic systems. Even this model requires the Moine sediments to be older than 1028Ma (with the Morar Pelite dates referred to below being diagenetic). However, the gneiss is an S-type granite (*sensu* Chappell & White 1974) whose composition is consistent with derivation by partial melting of Moine-like sediments (section 17.3.1) - it is therefore likely to have formed during high-grade crustal metamorphism, rather than 250Ma before any significant tectonothermal event. In addition, segregation veins in the gneiss (which from Aftalion & van Breemen's (1980) description belong to the $MP1_g$ set of sections 12.3 and 17.3.1) show similar isotopic features, and as they certainly cut the $S1_g$ fabric it must be at least as old as 1028Ma. Aftalion & van Breemen (op cit.) also obtained U/Pb monazite dates of 455±2Ma, and a 455±60Ma Rb/Sr isochron on 5 - 8cm wide slabs of metasediment. The former imply cooling through 530°C at 455Ma, the latter that solid diffusion was effective over 5 - 8cm in the interval from the beginning of Caledonian metamorphism until 455Ma (but not over 20cm or so - the size of Brook *et al.*'s (1976) samples).

In contrast, van Breemen *et al.* (1974) reported Rb/Sr dates or *715±5Ma from muscovite books in pre-D2 pegmatites in the high-grade Morar Division of Loch Eilt, partially-reset dates of *c.* *535Ma from a muscovite crenulated by D2, and *615-661Ma from muscovite in pre-D2 pegmatites at Glenfinnan. Thus the pegmatites must be at least *715Ma old, and D2 (of Brown *et al.* 1970. = $D2_g$) is probably post-535Ma. Rb/Sr muscovite dates on post-D2 pegmatites ranged from *436±30Ma to *446±50Ma at Loch Eilt, *432±6Ma to *440±8Ma in the granitic gneiss and *427±20Ma to *433±20Ma in the Glenfinnan Division, where a near-concordant U/Pb monazite date of 450±10Ma was also obtained. Brewer *et al.* (1979) combined the Glenfinnan Division results to obtain an Rb/Sr muscovite date of 430±8Ma, and a slightly discordant U/Pb monazite date of 429^{+11}_{-4} Ma. Thus the MP2 pegmatites were either emplaced at 430-440Ma or cooled through 500-530°C at that time, and D2 must be older than this. The discrepancy with the 455±2Ma monazite date from the granitic gneiss perhaps suggests that 430Ma represents intrusion rather than cooling. The MP1 pegmatites (as distinct from the migmatitic lits and the granitic gneiss's segregations) could be giving partially-reset Grenville ages or could date from a separate Morarian episode - the latter suggestion being supported by a U/Pb zircon date of 740±30Ma from Loch Eilt (van Breemen *et al.* 1978). K/Ar mica ages of *c.* 420Ma (Brewer *et al.* 1979) date cooling through 300-350°C.

Morar-Knoydart

It was in this area that two of the original Morarian dates were obtained, from the Knoydart and Sgurr Breac pegmatites (Fig. 14.7). Muscovite books from both pegmatites have given Rb/Sr dates in the range *745-765Ma ($\pm$15Ma) as well as *688$\pm$15Ma (Sgurr Breac) and 720$\pm$15Ma (Knoydart) - Giletti et al. (1961), Long & Lambert (1963). K/Ar mica dates of *467$\pm$20Ma and 454$\pm$20Ma were obtained from the Knoydart pegmatite by Miller & Brown (1965), and van Breemen et al. (1978) obtained concordant U/Pb monazite dates of *770$\pm$10Ma and *780$\pm$10Ma, while discordant zircons gave an upper intercept at 815$\pm$30Ma. This general agreement was taken to support a Morarian age for pegmatite emplacement - but since the U/Pb monazite and Rb/Sr muscovite systems have similar blocking temperatures, this is not a valid test. The younger muscovite dates could reflect some disturbance during Caledonian metamorphism (biotite-grade according to section 4.3), which at perhaps 450°C would have completely reset the K/Ar and Rb/Sr biotite systems. Fitch et al. (1969) reported an Ar^{39}/Ar^{40} plateau date of *758$\pm$10Ma for a muscovite from the Knoydart pegmatite. The agreement of all these ages and the concordance of the monazite age favour intrusion or cooling at c. 770Ma, rather than Grenville crystallisation and cooling followed by Caledonian isotopic disturbance. The zircon date could perhaps be explained by recent lead loss from discordant zircons which had intercepts at 1000Ma (crystallisation) and 450Ma (resetting) or 1700Ma (source) and 770Ma (crystallisation). Thus either a Morarian or a Grenville age could be supported, although in the latter case slow cooling would be indicated for the first 200 - 250Ma.

Brook et al. (1977) obtained an Rb/Sr whole-rock date of 1002$\pm$94Ma from samples of the Morar Schist at Druimindarroch (recalculation by Brewer et al. 1979). Unfortunately, the slope of the "errorchron" is critically dependent on two Rb-rich samples; if these were removed, a date of c. 750Ma would be obtained. Comparable whole-rock ages have been obtained from other Moine pelites - 769$\pm$50Ma for the Lochailort Pelite at Lochailort (Brewer et al. 1979), 720±120Ma for the Morar Schist at Glenborrodale and 700±100Ma for the Glenfinnan Division at Resipole (van Breemen et al. 1978). The last locality was assigned to the Morar Division by Johnstone (1975) but is here considered to lie in the Glenfinnan Division - cf. Chapter 6. The data points are widely scattered, and the last two "errorchrons" at least probably reflect Caledonian resetting - the Glenborrodale sample comes from a zone of very high late deformation (O'Brien 1981), while the Resipole sample lies within the zone of post-slide metamorphism and migmatisation discussed in Chapter 6. Samples collected by Brewer et al. (1979) from small areas at Lochailort gave Caledonian whole-rock ages (average 413±7Ma), as did samples from Knoydart (average 467±20Ma). The latter came from just above the Knoydart Slide, which may explain their resetting despite the preservation nearby of Precambrian Rb/Sr muscovite dates. Brewer et al. (1979) argued that these data could be combined to define a major isotopic event at 1004±28Ma, with the surprising corollary that all their Moine pelites were homogenised with respect to strontium isotopes (Sr^{87}/Sr^{86} =0.7092 ± 0.0007) at that time.

Rb/Sr biotite dates of c. 433Ma have been obtained from Knoydart (Brewer et al. 1979), while van Breemen et al. (1978) obtained dates of c. 450Ma and c. 420Ma from muscovite and biotite in western Morar. K/Ar mica dates range from 885Ma to *410±10Ma (Giletti et al. 1961, Miller & Brown 1965, Brewer et al. 1979), while Long & Lambert (1963) obtained Rb/Sr muscovite/whole-rock isochrons of *405-435Ma, except on the west coast where ages up to *570±15Ma were found. This indicates cooling of the eastern Morar Division at c. 410Ma, but suggests either earlier cooling or incomplete resetting in the west. Local anomalies may result from local mica recrystallisation and grain-size reduction caused by D3 and D4 deformation, although Moorbath (1969) also raised the possibility of excess argon causing anomalously old biotite dates in the west.

Thus there is strong evidence for a major Precambrian metamorphic event in this area - certainly older than 780Ma, and older than 885Ma if all the mineral ages are accepted: the evidence for this being a Grenville event is still not absolutely convincing, although none of the whole-rock data rule out such a model. Likewise the pegmatite evidence does not strongly favour a Morarian over a Grenville event, although in the latter case, slow cooling would appear to be a better explanation than partial resetting.

This Precambrian event must correspond to the MS2-MP2 metamorphism of Morar, or the MP1 metamorphism of the abbreviated eastern sequences (cf. Tables 14.1, 14.3) and by implication to that dated at Glenfinnan. D3 (Knoydart Slide) deformation is probably dated by the 467Ma whole-rock ages from the overlying pelite, and in any case it and the subsequent metamorphism must predate the 433Ma mineral ages. Brewer et al. (1979) used the younger date from the Lochailort Pelite (413Ma) to infer staggered uplift on first the Knoydart then the Sgurr Beag slide. However, the slide has been correlated with F2 of Brown et al. (1970), which is pre-430Ma, and it is folded by $F3_a$ near Salen, where micas (probably $MP3_a$) gave *418-428Ma cooling ages (Miller & Brown 1965). In addition, Rb/Sr muscovite dates of 438±10Ma and 439±10Ma were obtained by van Breemen et al. (1978) at Resipole, in the area of $MP3_a$ metamorphism and migmatisation. $MP3_a$ temperatures must have been above 500°C, so these dates provide a minimum age for $D3_a$; it is thus more likely that the younger dates in the east reflect the greater intensity of D4 (Morar sequence) deformation and metamorphism, and slower cooling.

Glen Dessary-Loch Quoich

The Glen Dessary syenite has recently been dated by van Breemen et al. (1979b). It is certainly folded by F3 (=$F3_q$) and possibly by F2 (=$F2_q$) but postdates migmatisation and isoclinal folding in the Moines. It produced sillimanite+K-Feldspar-bearing aureole assemblages, which were downgraded to muscovite+ quartz during Caledonian (MP2?) metamorphism, pyroxene in the intrusion being altered to hornblende. The best date for intrusion of the syenite is a 456±5Ma average Pb^{207}/Pb^{206} age from slightly discordant zircons. A U/Pb sphene result of 445±5Ma probably dates metamorphism of the syenite, or subsequent cooling through c. 550°C, while a 430±4Ma K/Ar hornblende age from the syenite dates cooling through 500°C. A 427±8Ma Rb/Sr muscovite date on a late pegmatite in the Moines is consistent with this, and could date intrusion of the pegmatite (cf. the MP2 pegmatites at Glenfinnan). K/Ar mica and Rb/Sr biotite ages of 403±8Ma to 412±7Ma date cooling through 300-350°C.

The Quoich granitic gneiss, although narrower and more highly-deformed, is "isotopically analogous to the granitic gneiss at Glenfinnan" (Piasecki & van Breemen 1979). Whole-rock samples give an Rb/Sr age of c. 1000Ma if Brook et al.'s (1976) initial Sr^{87}/Sr^{86} ratio is used, and U/Pb zircon results are similar to those of Pidgeon & Aftalion (1978).

The Quoich gneiss carries the $S1_q$ fabric of the adjacent Loch Eil Division and contains metasedimentary xenoliths (Chapter 11, section 17.3.1); like the Glenfinnan Division in Ardgour, the Loch Eil Division and its first fabric must be pre-Caledonian. No pegmatites have been dated in this area to confirm the Caledonian age of $D2_q$, but by analogy with Glen Dessary, $D3_q$ and possibly $D2_q$ are likely to be post-456Ma. $D3_q$ is probably pre-430Ma, given the pegmatite dates and the fact that the main metamorphism was $MP2_q$. The biotite-grade (> 400°C?) $D3_q$ event must predate the c. 410Ma mineral ages. Van Breemen et al. (1979b) suggested that this syenite was analogous to the Assynt alkaline intrusions, and that its formation was due to lithospheric loading consequent on thrusting - in that case the Sgurr Beag Slide (which defined a $D2_q$ deformation front) was probably the relevant thrust, and its movement spanned the 456Ma date of the intrusion.

Loch Monar-Carn Chuinneag

This is a classic area for Moine geochronology, containing as it does the pre-Caledonian Carn Chuinneag granite and the Morarian Carn Gorm pegmatite (Fig. 14.7), both being intrusive into the Morar Division. The Carn Gorm pegmatite predates local (Wilson 1975) D2, and has yielded Rb/Sr muscovite dates of c. *685Ma, *730±25Ma and c. *765Ma (van Breemen et al. 1974, Long & Lambert 1963, Pidgeon in Dunning 1972). Van Breemen et al. (1974) carried out a detailed study on a muscovite book and showed that the apparent age decreased outwards from *740Ma to *712Ma - they suggested that *740Ma was the closest approach to the age of emplacement. The Carn Chuinneag granite also predates local D2 (Shepherd 1973) and has been dated at *560±10Ma and 568±3Ma by Rb/Sr whole-rock (Long 1964) and U/Pb zircon methods (Pidgeon & Aftalion 1978). K/Ar mica dates in the granite and its aureole range from *421±19Ma to *432±19Ma (Giletti et al. 1961, Miller & Brown 1965). A post-Monar pegmatite has yielded a 438±6Ma Rb/Sr whole-rock date (van Breemen et al. 1974) and a syn- to post-Monar pegmatite a *421Ma K/Ar muscovite date (Brown et al. 1965).

The Carn Gorm pegmatite must have a true age of greater than 740Ma (either Morarian or Grenville), implying a Precambrian origin for S1 in this part of the Moines. The Carn Chuinneag granite confirms this, and also the Caledonian origin of D2. Wilson & Shepherd's (1979) D2 probably corresponds to the post-slide $D2_b$, $D3_s$ etc. (section 14.3); since development of the slide-age fabric is sporadic in the Morar Division, it is not possible to state whether the granite also predates the Sgurr Beag Slide. However, there is circumstantial evidence that this is so. The Carn Chuinneag aureole superimposes low-pressure cordierite+andalusite assemblages on the MP1 (Precambrian) garnet-grade mineralogy; "knotten" of biotite after cordierite replace regional garnet (Long & Lambert 1963) but are deformed by D2 (Wilson 1975). This implies P,T conditions of 500-650°C, <5kb, probably <3kb in the aureole (Wilson & Shepherd 1979). The MP2 metamorphism took place at much higher pressures (certainly >5kb), the relatively anhydrous cordierite+andalusite-bearing rocks of the aureole producing kyanite at temperatures where MP1 garnet+mica assemblages were still stable in the surrounding Moines (cf. Long & Lambert 1963). The most likely cause of this is the overthrusting of the Glenfinnan Division on the Sgurr Beag Slide, although a pressure increase of several kilobars implies that the Glenfinnan Nappe included higher rocks now removed by erosion.

A model which would be consistent with the metamorphic history of nearby parts of the slide follows. At 560Ma, the granite was intruded into the Moines (of Precambrian garnet grade) at perhaps 10km depth; cooling followed. At c. 460Ma, thrusting of the Glenfinnan Division on the Sgurr Beag Slide and loading of the Morar Division of the Carn Chuinneag area (to perhaps 20km) took place, but temperatures were too low for significant recrystallisation of the hornfelses in the absence of deformation. At c. 440 Ma, D2 (Shepherd) structures extended throughout the Moines, and the granite and its aureole were deformed; temperatures rose and metamorphism of the aureole assemblages occurred. Cooling to 300°C was completed by c. 420Ma. The post-Monar (=D2?) pegmatite with its 438Ma age would be consistent with this sequence.

Sutherland

On the whole, only Caledonian mineral dates have been obtained from this area, although it is likely that both Morar and Glenfinnan Divisions had a Precambrian history (sections 14.3, 9.5.3). Recently, late Precambrian (c. 650Ma) Rb/Sr whole-rock dates have been obtained from granitic members of the Strath Halladale complex (Lintern et al. in press). This, like the Loch Coire complex (Figs. 2.1, 14.7) consists of Glenfinnan Division migmatites and Lewisian inliers intruded by later igneous material - that at Loch Coire is generally assumed to be Caledonian, but these results suggest that it may in part be older.

K/Ar mineral ages in the range *400-430Ma have been obtained throughout the Morar Division and Sutherland migmatites (Brown et al. 1965, Miller & Brown 1965). The average age is about *420Ma, but there appears to be a slight tendency for the younger ages to occur in the vicinity of major intrusions or vein complexes, where cooling was probably slower (cf. *428Ma near Tongue with *395Ma NE of Lairg, near the Rogart granite). Giletti et al. (1961) reported a *433±9Ma Rb/Sr biotite date from Bettyhill. These are presumably minimum ages for Caledonian metamorphism. A *452±20Ma K/Ar muscovite date (Miller & Brown 1965) on a sample from Oykell Bridge showing strong D2 rodding may come closer to the true age of metamorphism - being in the west, close to the thrust belt, it should have cooled relatively rapidly.

Soper & Brown (1971) considered the Strath Vagastie suite of intrusions to predate D2, and drew the analogy with the Carn Chuinneag granite. However, it was argued in section 9.5.2 that convincing evidence for a pre-F2 age was lacking, that they were emplaced over a period of time, and that the simple shear fabric which they carry was superimposed on the F2 folds and could correspond to some later event in the thrust zone. Fergusson (1978) has also suggested a post-F2 age. Pidgeon & Aftalion (1978) derived a 405±11Ma age from a U/Pb zircon lower intercept. A sphene analysis lay on this discordia, close to the concordia (but not on it, as claimed by Pidgeon & Aftalion); it could be assigned a somewhat older age, perhaps 420-440Ma with recent lead loss (cf. the Strontian sample in the same paper). A post-F2 age allows the intrusions to be as late as 405Ma, but given the amount of ductile deformation they have suffered, an older date would be more geologically plausible.

The Loch Borrolan syenite (lying within the zone of duplexes beneath the Moine Thrust) has been dated at c. 430Ma (van Breemen et al. 1979a - U/Pb zircon, sphene); it is broadly synchronous with late (brittle) movement on the Moine Thrust. However, the K/Ar biotite date of Brown et al. (1968) shows that slow cooling or some additional disturbance continued intil *402±9Ma. Their older (417±10Ma) K/Ar date on the undeformed Ben Loyal syenite within the Moine Nappe suggests that continued movement in the thrust zone was a factor - perhaps deforming or burying the Loch Borrolan body.

Other areas

South of the Great Glen, Piasecki & van Breemen (1979) have reported Precambrian dates from "Old Moine" rocks (cf. section 2.1). The Ord Ban granitic gneiss is similar to that at Ardgour, and its Rb and Sr isotopes indicate a maximum age of c. 1300Ma (hence it cannot be a slice of Lewisian basement), but no clear Grenville date has been obtained. Pegmatites which were deformed in a slide zone gave Rb/Sr muscovite ages ranging from 573±13Ma to 718±19Ma, the latter being the least deformed. This invites comparison with the Caledonian deformation of Morarian pegmatites reported by van Breemen *et al.* (1974), but does not add to the discussion of whether the *c.* 750Ma ages relate to cooling or intrusion.

K/Ar mica dates of *413±18Ma and *437±19Ma have been obtained from the Tarskavaig "Moines", *429±19Ma from west of Strontian, *433Ma from the Ross of Mull and *440±19Ma from Glen Urquhart (Miller & Brown 1965). These are all typical Caledonian cooling ages, as are Rb/Sr biotite dates of *450±10Ma from Fannich and *435±50Ma from the Glenelg Lewisian.

14.4.3 Conclusions

The dates are summarised in Table 14.4. The reality of a Grenville event is confirmed in the Glenfinnan Division, and in that part of the Loch Eil Division which contains the granitic gneiss. The evidence in the Morar Division is not 100% convincing, but there are strong indications that it too was deformed and metamorphosed at *c.* 1000Ma rather than *c.* 750Ma. Similar dates have not yet been obtained from the Sutherland Moines, but structural correlations and outcrop continuity suggest that they are the same age. This event corresponds to D1 and D2 of western Morar and Knoydart, $D1_k$, $D1_s$ etc. (the pre-slide event in the Glenfinnan Division and northern Morar Division), and regional *lit-par-lit* migmatisation in the Glenfinnan Division, Knoydart Pelite and Loch Eilt Morar Division.

The "Morarian" pegmatites could have formed in this event, but being discrete intrusions which were not locally-derived, they need not be directly linked with the subsolidus regional migmatisation (although they could of course have been produced by partial melting at greater depth in the same metamorphic event). The two main alternatives are that they are Grenville, the *c.* 750Ma dates reflecting slow cooling or a Morarian thermal event (their concordance making partial Caledonian resetting unlikely), or that they form a distinct swarm unrelated to deformation or regional metamorphism. Recent work by D. Powell (verb. comm.) tends to support the latter interpretation, in that he has found Morarian pegmatites cutting low-grade Morar Division rocks west of Lochailort; in that case the oldest dates (c. 770Ma) should come closest to the age of intrusion. As was mentioned in section 14.3, the existence of two genuine Precambrian events can only be sustained if the structural correlations in Tables 14.1 and 14.3 are correct, i.e. if D2 in the eastern Glenfinnan Division is Caledonian, probably corresponding to D3 in the western Morar and westernmost Glenfinnan divisions, while D1 in the eastern Glenfinnan Division is a composite event embracing D1 and D2 of the less-deformed western areas. The Caledonian age of this eastern D2 is confirmed in the south by the spread of younger ages obtained from "Morarian" micas deformed in D2.

The Carn Chuinneag granite appears to have been intruded into an anorogenic environment, between the Precambrian and Caledonian events - it certainly provides a maximum date for Sutherland D2, and probably for the Sgurr Beag Slide as well. Other evidence for the age of the slide is provided by the Knoydart rocks which have suffered strong D3 (but little subsequent) deformation and biotite-grade metamorphism, and from the Glen Dessary syenite - c. 460Ma is the best estimate for sliding and the Glenfinnan Division D2. If the syenite is post-D2, these events could be somewhat older (post-560Ma, pre-460Ma). If the 455Ma monazite from Glenfinnan dates cooling after the amphibolite facies $MP2_g$ metamorphic peak, this provides a minimum age for $D2_g$ and perhaps a maximum for $D3_g$ - likewise the 445Ma sphene date from the Glen Dessary syenite.

Minimum ages are provided for $D3_q$, $D3_a$, D2 (Sutherland) etc. by later pegmatites at 427±8Ma and 438±6Ma; given the prevalence of 420-430Ma K/Ar dates in areas of $MP3_k$, $MP2_b$ etc. amphibolite facies metamorphism, an age of 440-450Ma would be reasonable, allowing some time for cooling. Van Breemen et al.'s (1978) Rb/Sr muscovite dates (438±10Ma) from Resipole, in an area of $MP3_a$ migmatisation, are consistent with such an interpretation, and the Oykell Bridge sample would agree within errors. The Glenfinnan pegmatites are somewhat anomalous in that case (rather young to be pre-D3) - they may have yielded cooling dates, or perhaps their structural dating is in error. Late movements on the Moine Thrust Zone are dated at 430Ma, and the main ductile movement must clearly have been earlier; metamorphism, and presumably the later phases of deformation which were not mapped, continued until 410-420Ma, by which time the Newer Granites were being intruded.

Thus the main time-markers in the Moines are a c. 1000Ma orogeny with extensive deformation and migmatisation, a possible 770Ma intrusive event involving Morarian pegmatites, intrusion of the Carn Chuinneag granite at 560Ma, movement on the Sgurr Beag Slide and intense deformation to the east of it at c. 460Ma, and ductile movement in the Moine Thrust Zone and the main phase of Caledonian folding at 440-450Ma. The last two events probably succeeded one another more-or-less continuously. Brittle movement in the thrust zone continued beyond 430Ma, but all regional folding in the Moines predates the 435±10Ma Strontian and 423±4Ma Ross of Mull granites (Pidgeon & Aftalion 1978, Beckinsdale & Obradovitch 1973). Late movements are attested to by the simple shear deformation of suitably-oriented microdiorites in the SW Moines (Smith 1979).

Caledonian deformation and metamorphism in the Moines is distinctly younger than that in the Dalradian (490-500Ma - Bradbury et al. 1976) and may well have been in response to the crustal thickening which this produced to the southeast (e.g. Wells & Richardson (1979) suggested a peak crustal thickness in excess of 60-70km). The ages of the Sgurr Beag Slide and the Moine Thrust are consistent with such an interpretation, deformation migrating towards the foreland as the thickened pile collapsed, as is the evidence of Coward (1982) that gravity spreading was an important driving mechanism in the Moine Thrust Zone.

The evidence for a Grenville event affecting all the Northern Highland Moines tends not to favour correlation with the Torridonian of the foreland. Moorbath (1969) obtained supposed diagenetic Rb/Sr whole-rock dates of *970±25Ma (Lower Torridonian) and *790±20Ma (Upper Torridonian) from red shales within the sequence. The former date is sufficiently close to allow equivalence if it represents a later disturbance or if closure due to diagenesis was delayed by

50Ma or so. However, J.D.A. Piper (paper read at Joint Association for Geophysics meeting, September 1982) has suggested that the palaeomagnetic poles for both Lower and Upper Torridonian are inconsistent with a 1000-800Ma age, but fit well with Keewanawan poles from Greenland and Canada (Lower Torridonian 1100Ma, Upper Torridonian 1040Ma), so perhaps the Rb/Sr dates do reflect later disturbance. *

* These results have recently been published as: Smith, R.L., Stearn, J.E.F. and Piper, J.D.A. (1983). Palaeomagnetic studies of the Torridonian sediments, N.W. Scotland. Scott. J. Geol. 19, 29-45.

CHAPTER 15

Geothermometry and geobarometry

Within the Moines, index minerals for the higher grades of metamorphism (staurolite, kyanite and sillimanite) are particularly rare. Temperatures and pressures, estimated from the compositions of minerals in equilibrium assemblages, can help to define variations in grade from place to place, and can also be compared with experimental work on possible melting reactions in metasedimentary migmatites.

In general, thermodynamic data for minerals are presented in calories for ΔH, e.u. (entropy units = $calK^{-1}$) for ΔS and $calbar^{-1}$ for ΔV (1 $calbar^{-1}$ = $41.84cm^3$). Temperature is in K (degrees Kelvin) and pressure in bars. Final results were converted to °C and kilobars (kb).

15.1.1 General principles

The method of estimating pressure and temperature conditions of formation basically involves application of the Van't Hoff isotherm (e.g. Wood & Fraser 1977, p.66 et seq.). Consider a reaction involving w moles of component A, x moles of component B, y moles of C and z moles of D, as below, where a, b, c and d are the phases in which these components reside.

$$wA + xB = yC + zD$$

phase a b c d

At equilibrium, $w\mu_A^a + x\mu_B^b = y\mu_C^c + z\mu_D^d$

(where μ_A^a is the chemical potential of component A in phase a, etc.) Defining activity of component A in phase a in relation to a standard state of pure A:

$$w\mu_A^a = w\mu_A^o + wRT\ln a_A^a$$

(where μ_A^o is the chemical potential of pure A and a_A^a is the activity of component A in phase a)

This gives the equation:

$$w\mu_A^{\circ} + wRT\ln a_A^a + x\mu_B^{\circ} + xRT\ln a_B^b = y\mu_C^{\circ} + yRT\ln a_C^c + z\mu_D^{\circ} + zRT\ln a_D^d$$

Defining $\Delta G^{\circ} = y\mu_C^{\circ} + z\mu_D^{\circ} - w\mu_A^{\circ} - x\mu_B^{\circ}$, the above equation can be rearranged to:

$$\Delta G^{\circ} = -RT\ln \left[\frac{\left(a_C^c\right)^y \cdot \left(a_D^d\right)^z}{\left(a_A^a\right)^w \cdot \left(a_B^b\right)^x}\right]$$

The term in brackets is referred to as the <u>equilibrium constant</u>,K.

Taking the standard state of each component to be the pure phase at the pressure and temperature of interest, μ° can be replaced by the molar free energy at P and T, i.e. $\mu_A^{\circ} = G_A^{\circ}$, etc. $\Delta G^{\circ}_{P,T}$ can then be calculated from tables of entropy, enthalpy and heat capacity. Given mineral compositions from microprobe analyses (section 2.3), it is possible to define a line in P,T space (curve of given K value) at which all the components could have been in equilibrium. The significance of this line depends on three factors:

i) attainment (and preservation) of equilibrium between the relevant phases;

ii) relationship of this equilibration to some geological event;

iii) accuracy of the thermodynamic data for the pure phases and for any activity coefficients, or of an empirical calibration.

For reactions involving fluids, it is convenient to use standard states of P,T for the solid components and 1 bar, T for fluids. There is therefore an additional activity term for the fluid, which is best kept separate from those which relate to solid solution in the mineral phases. The Van't Hoff isotherm can then be rewritten:

$$\Delta G^{\circ} = -nRT\ln a_{fl} - RT\ln K$$

where n is the number of moles of fluid participating in the reaction. Using 1 bar, T as the standard state for fluids one obtains, e.g. for H_2O:

$$RT\ln a_{H_2O} = RT\ln\left[\frac{f_{H_2O}}{f^{\circ}_{H_2O}}\right] = RT\ln\left[\frac{P\Gamma_{H_2O}}{P^{\circ}\Gamma^{\circ}_{H_2O}}\right]$$

where f_{H_2O} is water fugacity, Γ_{H_2O} the fugacity coefficient, and f°, P° and Γ° the appropriate standard state values. At 1 bar, T>200°C, $\Gamma_{H_2O} \simeq 1.0$, so that $a_{H_2O} = f_{H_2O} = P\Gamma_{H_2O}$.

Values of Γ_{H_2O} can be obtained from tables (e.g. Wood & Fraser 1977, p. 281).

ΔG° may be calculated from data in Helgeson et al. (1978, tables 8 & 9). These give ΔH°_f and S° at 1 bar, 298K and heat capacity data at geological temperatures. The relevant equation is:
$\Delta G^\circ = \Delta H^\circ_{1,T} - T\Delta S^\circ_T + (P-1)\Delta V^\circ_s$.
For the hypothetical reaction discussed above,

$$\Delta H^\circ = y(\Delta H^\circ_f)_C + z(\Delta H^\circ_f)_D - w(\Delta H^\circ_f)_A - x(\Delta H^\circ_f)_B$$

ΔS° and ΔV°_s are obtained in a similar fashion.

In practice, entropy and enthalpy data are usually only available for 1 bar, 298K, but they can be calculated, for each pure phase at given T, using heat capacity data (the effect on ΔV_s due to thermal expansion can be neglected). Using the equations in Wood & Fraser (op.cit.) pp. 22 - 29:

$$\int_{298}^{T} dH = \int_{298}^{T} C_p dT \quad \text{i.e.} \quad H_T - H_{298} = \int_{298}^{T} C_p dT$$

(where C_p is the heat capacity)

C_p is usually represented by the expression $C_p = a + bT + c/T^2$.
This integrates to give the expression:

$$H_T = H_{298} + \left[aT + \frac{bT^2}{2} - \frac{c}{T} \right]_{298}^{T}$$

In practice, Δa, Δb and Δc for the reaction are calculated in the usual manner, giving $\Delta Cp = \Delta a + \Delta bT + \Delta c/T^2$, and

$$\Delta H_T = \Delta H_{298} + \left[\Delta a.T + \frac{\Delta b.T^2}{2} - \frac{\Delta c}{T} \right]_{298}^{T}$$

Similarly, for entropy, $dS = \dfrac{C_p dT}{T}$

Integrating, $S_T - S_{298} = \int_{298}^{T}\left(\frac{a}{T} + b + \frac{c}{T}3\right) dT$

and $\Delta S_T = \Delta S_{298} + \left[\Delta a.\ln T + \Delta b.T - \frac{\Delta c}{2T}2\right]_{298}^{T}$

In a solid-solid reaction, Δa, Δb and Δc tend to cancel out, i.e. $\Delta C_p=0$, so this correction can be ignored. For reactions involving fluids, the correction is small over a range of 100 - 200°C, so (say) ΔH_{800} could be used from 700 - 900K.

The final equation used is therefore:

$$\Delta G° = \Delta H^{\circ}_{1,T} - T\Delta S^{\circ}_{T} + (P - 1)\Delta V^{\circ}_{s} = -nRT\ln f_{H_2O} - RT\ln K$$

Examples of its use, and rearrangement to solve specific problems, are given in the succeeding sections.

In calculating K, the simplest method is to assume ideal solution. In this case, $a^a_A = \gamma^a_A . X^a_A = X^a_A$, since γ^a_A (the activity coefficient for component A in phase a) is unity. X^a_A is determined on a statistical basis. For example, in alkali feldspar, $KAlSi_3O_8 \leftrightharpoons NaAlSi_3O_8$, a^{fs}_{Na} equals the mole fraction of albite. In garnet, since there are three sites, the mole fraction is cubed, i.e. in $Mg_3Al_2Si_3O_{12} \leftrightharpoons Fe_3Al_2Si_3O_{12}$,

$a^{ga}_{py} = \left(\frac{Mg}{Mg+Fe}\right)^3$. The major exception to this rule is plagioclase, where $NaAlSi_3O_8 \leftrightharpoons CaAl_2Si_2O_8$ involves exchanging two atoms, so a^{pl}_{alb} should be given by $\left(\frac{Na}{Na+Ca}\right) . \left(\frac{Al}{Al+Si}\right)$ (the latter on the AlSi site ,which is available for exchange). However, $a^{pl}_{alb} = \left(\frac{Na}{Na+Ca}\right)$ agrees better with experimental work. This is probably due to preservation of local charge balance, i.e. allowing Na/Ca and Al/Si to behave independently would give rise to local $NaAl_2Si_2O_8^-$ and $CaAlSi_3O_8^+$ molecules.

In practice, most solid solutions show some deviation from ideality. This may be expressed in terms of excess functions (e.g. $\bar{G}_{xs}$, $\bar{V}_{xs}$, $\bar{S}_{xs}$), describing the amount by which a given thermodynamic property of the solution differs from that of a hypothetical ideal solution of the same bulk composition (Wood & Fraser 1977, p.101 et seq.; Thompson 1967). It is then possible to determine activity coefficients, which define the activity-composition relationships of the solution. Taking P, T standard states, as before,

$$\mu_A^a = \mu_A^o + RT\ln a_A^a = \mu_A^o + RT\ln(X_A^a.\gamma_A^a) \quad \text{and} \quad (\mu_A^a)_{xs} = RT\ln\gamma_A^a$$

for a binary solution of components A and B in phase a.

The relationships of G_{xs} to P,T and X, and of γ to G_{xs} and X, have been determined for a number of solid solutions, using various mixing models (e.g. see Wood & Fraser 1977, p.113). In general, $\gamma_A^a \rightarrow 1.0$ as $X_A^a \rightarrow 1.0$ and γ_A^a approaches a constant as $X_A^a \rightarrow 0.0$ (Henry's Law). Positive deviations from ideality ($\overline{G}_{xs} > 0$) reflect a tendency towards unmixing, with development of a solvus or miscibility gap, and are the general rule in silicate solid solutions; these are associated with activity coefficients greater than unity. Various mixing models define $\overline{G}_{xs}$ in terms of interaction parameters, W, and composition. The models used in this chapter are summarised below.

Quartz, kyanite, sillimanite There were taken to be pure phases, with a = X = 1.0.

K-feldspar This was treated as an ideal solution, since all analyses used had $X_{Or}^{Kfs} \geqslant 0.9$; where K-feldspar was identified petrographically, but not probed, a composition of $X_{Or}^{Kfs} = 0.9$ was assumed.

Epidote-clinozoisite In the absence of a-X data, this was treated as an ideal solution, although given the size difference between Al^{3+} and Fe^{3+}, it is likely that $\gamma_{cz}^{ep} > 1.0$. However, most of the analysed minerals had $X_{cz}^{ep} > 0.5$, so the error should not be too large. Fe^{3+} was distributed between the M1 and M3 sites, assuming complete disorder within each site; a ratio of $X_{Fe}^{M1} : X_{Fe}^{M3} = 0.15 : 0.85$ was used (cf. Helgeson et al. 1978, p.187).

Plagioclase A value of $\gamma_{an}^{pl} = 1.28$ was taken for sodic plagioclase (Orville 1972); this was determined at 700°C, and may be an under-estimate of γ for very albitic compositions at lower temperatures. $\gamma_{an}^{pl} = 1.0$ was used for bytownite-anorthite compositions, and $\gamma_{ab}^{pl} = 1.0$ for all plagioclases; equilibria involving albite were not applied to very calcic plagioclases.

Garnet A quaternary regular solution model was applied (Ganguly & Kennedy 1974), using the W values in Table 15.1 (suggested by Ganguly 1979). This assumes that the magnitudes of $\bar{G}_{xs}$ etc. are symmetrical with respect to composition (and in this case, independent of P and T). The pyrope-grossular and almandine-grossular joins are probably asymmetric (Ganguly 1979, p.1023; Newton et al. 1977), but the symmetric model fits experimental data, at least to $X_{Ca}^{ga} = 0.3$. The equations are:

$$\bar{G}_{xs} = X_G X_A W_{AG} + X_G X_P W_{PG} + X_A X_P W_{AP} + X_G X_S W_{GS} + X_A X_S W_{AS} + X_P X_S W_{PS}$$

$$RT\ln\gamma_{gr}^{ga} = X_A^2 . W_{AG} + X_P^2 . W_{PG} + X_S^2 . W_{GS} + X_A X_P (W_{AG} + W_{PG} - W_{AP})$$

$$+ X_A X_S (W_{AG} + W_{GS} - W_{AS}) + X_P X_S (W_{PG} + W_{GS} - W_{PS})$$

(Wood & Fraser 1977, p.109).

Note that these values are per mole of divalent cation; the garnet formula used in the succeeding sections is $Ca_3Al_2Si_3O_{12}$, so $a_{gr}^{ga} = \left(X_{gr}^{ga} . \gamma_{gr}^{ga}\right)^3$.

White mica For the muscovite - paragonite solid solution, Eugster et al. (1972) have determined the following asymetric Margules expression (cf. Thompson 1967, pp.351-357):

$$\bar{G}_{xs} = (3082 + 0.170T + 0.082P) X_{pa} . X_{mu}^2$$

$$+ (4164 + 0.395T + 0.126P) X_{mu} . X_{pa}^2$$

where the terms in brackets are $(W_G)_{pa}$ and $(W_G)_{mu}$ respectively.

Activity coefficients are determined by:

$$RT\ln\gamma_{mu}^{wm} = (X_{pa}^{wm})^2 . (W_{mu} + 2(W_{pa} - W_{mu}) X_{mu}^{wm})$$

$$RT\ln\gamma_{pa}^{wm} = (X_{mu}^{wm})^2 . (W_{pa} + 2(W_{mu} - W_{pa}) X_{pa}^{wm})$$

In practice, the choice of P and T within $\pm$ 100°C and $\pm$2kb have negligible effects on the activity coefficients. The values obtained are per mole of alkali atoms. To calculate $X^{wm}_{pa_2}$ and X^{wm}_{mu}, ideal mixing on sites was assumed, i.e. $X^{wm}_{pa} = (X^{A}_{Na}).(X^{oct}_{Al})^2$. The analysed micas contained significant octahedral Mg, Fe and Ti, so that $X^{wm}_{pa} + X^{wm}_{mu} \simeq 0.75$. No data is available for this substitution, but it is probably associated with positive deviations from ideality, and the activity coefficients derived above can be regarded as minimum estimates.

Biotite Phlogopite-annite can probably be treated as an ideal solution (Wones 1972), but Moine biotites deviate strongly from this series. Biotite only appears in the garnet-biotite and Ghent & Stout (1981) equilibria (sections 15.1.2, 15.1.4). In both cases, empirical calibrations based on natural assemblages were applied, so as long as the Moine biotites have similar compositions, any activity coefficients will have been taken into account.

15.1.2 Geothermometry

The garnet-biotite geothermometer was used, that of Thompson (1976) proving the most appropriate.

The reaction is:

$$\underset{\text{garnet}}{Mg_3Al_2Si_3O_{12}} + \underset{\text{biotite}}{KFe_3AlSi_3O_{10}(OH)_2} = \underset{\text{garnet}}{Fe_3Al_2Si_3O_{12}} + \underset{\text{biotite}}{KMg_3AlSi_3O_{10}(OH)_2}$$

$$K = \frac{\left(a^{ga}_{alm}\right) \cdot \left(a^{bi}_{phl}\right)}{\left(a^{ga}_{py}\right) \cdot \left(a^{bi}_{ann}\right)}$$

Assuming ideal solution, or at least that activity coefficients cancel out, K reduces to $((Mg/Fe)_{bi}/(Mg/Fe)_{ga})^3$

Most workers find it convenient to use:

$K_D = (Mg/Fe)_{bi}/(Mg/Fe)_{ga}$, i.e. $K = K_D^3$

At equilibrium, $\Delta H^{o} - T\Delta S^{o} + (P-1)\Delta V + RT.\ln K_D^3 = 0$

i.e. $\Delta H^{o} - T\Delta S^{o} + P\Delta V + 3RT.\ln K_D = 0$

((P-1) can be approximated by P when the pressure is several k bars).

Since this is a solid-solid reaction, ΔH and ΔS will not change significantly with T, so at constant pressure, $\ln K_D = \frac{\Delta S^{o}}{3R} - \frac{(\Delta H^{o} + P\Delta V)}{3RT}$,

i.e. $\ln K_D$ is inversely proportional to T. In fact ΔV for this reaction is so small (-0.057 $calbar^{-1}$) that the effect of pressure is also minimal.

Three calibrations are available, each utilising different approaches.

Thompson (1976) used garnet-biotite data for four areas where the temperature was well constrained by a net of other reactions, and obtained a regression line through them (Fig. 15.1). Samples were chosen which appeared to be in equilibrium and lacked any sign of retrograde zoning; the temperatures obtained can be directly compared with Thompson's petrogenetic grids (Thompson 1974, 1976; Thompson & Algor 1977). Tracy (1978) found good agreement between Thompson's geothermometer and the grid of melting reactions referred to in section 15.3.1. Also, since natural garnets from pelites, containing Ca and Mn, and biotites containing octahedral Al were used, any activity coefficient due to these components should cancel out, if the Moine garnets have similar compositions.

Ferry & Spear (1978) performed experiments on synthetic pyrope-almandine and annite-phlogopite at 2070 bars. They calculated ΔH, ΔS from a regression line of $\ln K_D$ against 1/T, using the equation given above. Their results were:

$\Delta H = -12454$ cal, $\Delta S = -4.662$ e.u. This provides the equation:

$$4.662T - 12454 - 0.057P + 3RT\ln K_D = 0$$

2σ ranges for ΔH and ΔS were -8166 to -16784 cal and -0.143 to -9.277 e.u.. Fig. 15.1 shows the curve at 2.07 kb, and recalculated to 5kb for application to the Moines. The slope of lines of equal K_D can be readily calculated from the above equation by: $\frac{dT}{dP} = \frac{0.057}{4.662+3R\ln K_D}$. This gives 5-6°C/kb for $\ln K_D$ = 2.5-1.5. Since most results lie in the 5-8 kb range, this pressure correction can safely be ignored.

Thompson's calibration can be subjected to a similar thermodynamic analysis if a pressure is assumed. Since the rocks crystallised in the staurolite to garnet-cordierite zones, pressures of 4-6 kb would seem reasonable. 4 kb gives ΔH = -16290 cal, ΔS = -9.461 e.u.; 6 kb gives the same ΔS, but ΔH is -16176 cal. These values lie at the limit of Ferry & Spear's 2σ range; dT/dP ranges from 3.5-4.0°C/kb. Disadvantages of Ferry & Spear's calibration are that the minerals studied did not contain the impurities which are present in natural biotites and garnets, so activity coefficients may be required, and most of their data points are at high temperatures (>650°C).

Goldman & Albee (1977) used analyses from the literature, and oxygen isotope temperatures, to produce a plot of $\ln K_D$ vs 1/T. (Fig. 15.1). They also performed a statistical analysis to assign empirical corrections for Mn, Ca in garnet, and Ti, Al^{VI} in biotite. Their temperatures are much lower than those obtained using the other two calibrations, possibly due to the oxygen isotopes re-equilibrating during cooling.

For the first few rocks studied, all three methods were used in an attempt to determine their relative usefulness. The results are compared in Table 15.2.

In Glenfinnan Division rocks, Mn and Ca in garnet are generally fairly low, so there is little point in performing the laborious calculations involved in Goldman and Albee's thermometer. Their straightforward lnK *vs.* 1/T plot gives unacceptably low results - e.g. 450°C for garnet grade rocks from Mallaig, 550°C for sillimanite zone migmatites from Acharacle. The Morar Division rocks have high Mn and Ca in garnet (up to 50% of the garnet molecule), so it was hoped that Goldman & Albee's equations could be used. However, they gave very erratic results, so were not applied to any further rocks (Table 15.2). Thompson's curve was preferred to Ferry & Spear's, since it gave results which were sensible in terms of the mineral parageneses and the petrogenetic grids mentioned above. They also fitted well with curves such as kyanite = sillimanite, and muscovite + quartz = K-feldspar + sillimanite, calculated from Helgeson *et al.*'s (1978) data. These last also agreed with the grids of Thompson and co-workers. Ferry & Spear's calibration gives very high temperatures for the higher grade rocks (e.g. up to 700°C for rocks showing no sign of muscovite breakdown - cf. Tables 15.2, 15.4 - 15.7).

Use of Thompson's geothermometer and curves calculated from Helgeson *et al.*'s data gives a self-consistent set of results in which the relative positions of different areas should remain constant, even if the absolute P,T values later prove to be wrong.

A common problem in garnet-biotite geothermometry is the presence of zoning in garnet. In principle, only the edge of the garnet can be in equilibrium with matrix biotite. Since diffusion in garnets must be many times slower than in biotites (witness the common preservation of zoning in garnets), the garnet can be considered to extract material from a reservoir consisting basically of biotite. Since garnet is always more Fe-rich than coexisting biotite (cf. Fig. 15.1), Mg/Mg+Fe in the reservoir will increase as garnet grows, so biotite which was in equilibrium with the garnet core should have been more Fe-rich than the present matrix biotite. The net effect of this is to make garnet core/matrix biotite temperatures too low, if the garnet grew at constant temperature. This problem can be partially circumvented by choosing a rock with a high biotite/garnet ratio, so that the biotite approximates to an infinite reservoir. However, this usually means that the garnet is very Mn-rich (since most of the Mn in the rock will be in the garnet), so the temperatures obtained may be less reliable. A common feature is a narrow Mg-poor rim to the garnets. This could be due to exchange between biotite and garnet during cooling, or even later prograde metamorphism, or to growth of a thin shell of new garnet at lower temperatures. In either case, the next-to-rim compositions should give the best estimate of the metamorphic peak, since the matrix composition could not have changed much due to the growth of a very thin rim (remembering that a thin outer shell represents a relatively large volume of garnet). The question of zoning, and structural dating of individual garnets, is discussed on a rock-by-rock basis in section 15.2.

Some garnets in the high-grade areas are unzoned; this has been attributed by MacQueen & Powell (1977) to Caledonian homogenisation of originally-zoned Precambrian garnets. However, zoned and unzoned garnets exist side-by-side in rocks of slightly differing composition (e.g. at Kinlochourn), and in some cases (e.g. Inverie, Kinlochourn), early garnets are zoned but later ones unzoned. This suggests that the presence or absence of zoning is at least in part a primary feature. Zoning in trace elements, e.g. Mn, is produced at constant K_D if the refractory garnet depletes the matrix in this element (cf. fractional crystallisation from a melt). At higher temperatures, K_D is reduced, so garnets which grew at high temperatures should show less Mn zoning than those which grew at low temperatures. In addition, the effective composition of the reservoir may change to counteract the depletion effect, e.g. if the consumption of ilimenite releases Mn+Fe, or zoisite Ca. In terms of major components, a garnet which grew during rising or falling temperatures could be more or less strongly zoned than one which grew at constant temperature, depending on the T-X relations of the relevant divariant or multivariant reactions (cf. Tracy _et al_. 1976, Tracy 1978). All these mechanisms provide for control of growth zoning by temperature or bulk composition (which defines the garnet-forming reactions involved), and so are favoured by the relations described above. Indirect support is also provided by the fact that changes in garnet-biotite temperatures could be detected across the Sgurr Beag Slide, even in areas where the peak of Caledonian metamorphism was post-sliding, and Caledonian metamorphic zones cross the slide (e.g. Kinlochourn - cf. Tanner 1976).

15.1.3 Geobarometry

The most useful reactions for geobarometry are those with a large ΔV (since $dP/dT = \Delta S/\Delta V$); in practice, this means reactions involving grossular $\rightleftharpoons$ anorthite. The most commonly used reaction is:

$$\underset{\text{garnet}}{Ca_3Al_2Si_3O_{12}} + \underset{\text{kyanite}}{2Al_2SiO_5} + \underset{\text{quartz}}{SiO_2} = \underset{\text{plagioclase}}{3CaAl_2Si_2O_8}$$

Using data from Helgeson et al. (1978) and taking P,T standard states:

$\Delta H =$ +11525 cal ΔS + +36.55 e.u. ΔV = +1.5823 cal/bar

$$K = \frac{\left(a^{pl}_{an}\right)^3}{\left(a^{ga}_{gr}\right)} = \left(\frac{(Ca/Ca+Na+K)\ pl}{(Ca/Ca+Mn+Mg+Fe)ga}\right)^3 \cdot \left(\frac{\gamma^{pl}_{an}}{\gamma^{ga}_{gr}}\right)^3$$

(activity coefficients defined in section 15.1.1, quartz and kyanite pure phases).

$(Ca/Ca+Mn+Mg+Fe)_{ga}$ and $(Ca/Ca+Na+K)_{pl}$ will be referred to as X^{ga}_{Ca} and X^{pl}_{Ca} respectively.

Then: $\Delta H - T\Delta S + (P-1)\Delta V = RT\ln K$

$$\text{i.e.}\quad 11525 - 36.55T + 1.5823(P-1) = -3RT\ln\left(\frac{X^{pl}_{CA}}{X^{ga}_{Ca}}\right)\left(\frac{\gamma^{pl}_{an}}{\gamma^{ga}_{gr}}\right)$$

Unfortunately, kyanite- or sillimanite-bearing assemblages are rare in the Moines. In the absence of aluminosilicate, the line calculated from the above equation would give a maximum pressure. This calculation was performed for several rocks, but the pressures obtained (> 10kb) were too high to provide a useful constraint.

This reaction can be modified by coupling it with the paragonite-quartz breakdown reaction. The resulting equation describes the equilibrium between plagioclase, white mica and garnet:

$$\underset{\text{garnet}}{Ca_3Al_2Si_3O_{12}} + \underset{\text{white mica}}{2NaAl_3Si_3O_{10}(OH)_2} + \underset{\text{quartz}}{3SiO_2} = \underset{\text{plagioclase}}{3CaAl_2Si_2O_8}$$

$$+ \underset{\text{plagioclase}}{2NaAlSi_3O_8} + \underset{\text{fluid}}{2H_2O}$$ (abbreviated "the Na reaction")

$\Delta H_{1,298}$ = +52005 cal, ΔS_{298} = +113.218 e.u., ΔV_s = +1.05368 cal/bar

These were recalculated to 750K, using the heat capacity data in Helgeson et al. (1978) (cf. section 15.1.1):

$\Delta H_{1,750}$ = +57700 cal ΔS_{750} = + 121.389 e.u.

The equilibrium constant for this reaction is:

$$K = \frac{\left(a^{pl}_{an}\right)^3 \cdot \left(a^{pl}_{ab}\right)^2 \cdot \left(a^{fl}_{H_2O}\right)^2}{\left(a^{ga}_{gr}\right)^3 \cdot \left(a^{wm}_{pa}\right)^2 \cdot \left(a^{qtz}_{SiO_2}\right)^3}$$

Using the activity coefficients from section 15.1.1, with standard states of P, T for solids, and 1 bar, T for H_2O, $X^{fl}_{H_2O}$ = 1.0, the the following equation holds at equilibrium:

$$\Delta G^{\circ}_{P,T} = \Delta H^{\circ}_{1,T} - T\Delta S^{\circ}_{T} + (P-1)\Delta V_s + 2RT\ln f_{H_2O} = -RT\ln K$$

$$= -RT\ln \frac{\left(X^{pl}_{Ca}\right)^3 \cdot \left(X^{pl}_{Na}\right)^2}{\left(X^{ga}_{Ca}\right)^3 \cdot \left(X^{wm}_{NaAl}\right)^2} \cdot \frac{\left(\gamma^{pl}_{an}\right)^3 \cdot \left(\gamma^{pl}_{ab}\right)^2}{\left(\gamma^{ga}_{gr}\right)^3 \cdot \left(\gamma^{wm}_{pa}\right)^2}$$

A convenient form for computation is:

$$\frac{57700}{RT} - \frac{121.389}{R} + \frac{1.05368\ (P-1)}{RT} + \ln K = -2\ln f_{H_2O}$$

Since f_{H_2O} does not vary linearly with P, this equation can be solved by an iterative method (Wood & Fraser 1976, p.73), using the table of Γ_{H_2O} vs. T and P quoted in Wood & Fraser (op.cit.) p.281. In practice, a template was prepared with curves for K = 2 to K = 16384 (Fig. 15.2 (b)), and K values for particular rocks were interpolated. Two additional sources of error exist in this method: $X^{fl}_{H_2O}$ may be less than 1.0, and the activity coefficient for paragonite in white mica may be in error. In normal pelites at moderate grade, with no partial melting, the fluid is probably mostly water; the effects of low water pressure can be assessed qualitatively by incorporating a correction in K - e.g. $X^{fl}_{H_2O}$ = 0.7 would halve the value of K, resulting in a pressure increase of 0.5kb. From section 15.1.1, γ^{wm}_{pa} is likely to have been underestimated. If it is assumed that the muscovite-phengite-margarite series show similar departures from ideality to the muscovite-paragonite series, the probable error can be calculated. Typically, $X^{wm}_{mu} = 0.7$, $X^{wm}_{pa} = 0.05$; if X^{wm}_{mu} was 0.95, RTln $^{wm}_{pa}$ would increase by a factor of $(0.95/0.7)^2 = 1.84$, and K by a factor of $(1.84)^2 = 3.4$. From Fig. 15.2, this would result in a 1kb increase in pressure.

An alternative equilibrium is:

$$\underset{\text{garnet}}{Ca_3Al_2Si_3O_{12}} + \underset{\text{white mica}}{2KAl_3Si_3O_{10}(OH)_2} + \underset{\text{qtz}}{3SiO_2} = \underset{\text{K-feldspar}}{2KAlSi_3O_8} +$$

$$\underset{\text{plagioclase}}{3CaAl_2Si_2O_8} + \underset{\text{fluid}}{2H_2O}$$ (abbreviated "the Kfs reaction")

For this reaction (Helgeson *et al*. 1978)

$\Delta H_{1,298}$ = +53879 cal ΔS_{298} = +111.658 e.u. ΔV_s = +1.0833 cal/bar

Using heat capacity data, as before:

$\Delta H_{1,800}$ = +60561 cal ΔS_{800} = +127.679 e.u.

In principle, this reaction is a better choice, since the phases are more nearly pure, and activity coefficients should be close to unity (in the absence of K-feldspar, a minimum pressure is obtained). However, it gives unrealistically high pressures (10 -12 kb for most rocks, some > 15kb; Kfs minimum pressure is usually much greater than the Na pressure). This is apparently due to the heat capacity values suggested by Helgeson *et al*. (op.cit.) for K-feldspar. Using 1,298 values for ΔH and ΔS gives results which are compatible with the Na reaction.

An alternative method of calculation is given by Powell (1978 p.253). The need for a temperature correction to ΔH and ΔS arises mainly from the differing heat capacity of water in a fluid and structurally bound in a solid. The difference is reasonably constant for different minerals, so an empirical factor can be applied. Powell also includes linearised water fugacity data in this factor, so that at given T, P can generally be obtained without iteration. The equation is:

$$\Delta G^\circ = \Delta H^\circ_{1,298} - T\Delta S^\circ_{298} + P\Delta V_s + nF_{H_2O} + mF_{CO_2}$$

for a reaction involving n moles of H_2O and m moles of CO_2 on the right hand side. F_{H2O} is tabulated as coefficients a' and b', where F = a' + b'T, in Powell (1978) p.257 (here, ΔH is in kJ, ΔS in kJK^{-1}, T in K, and ΔV in kJ/kb (1kJ/kb = $10cm^3$)). F_{H_2O} values apply over 100°C and 2kb, so if the initial pressure estimate is near enough, only one iteration is needed. The more precise data of Helgeson, et al. (1978) (converted using 1cal = 4.184J) was preferred to that presented by Powell (1978, p.254).

Curves for both the Na and Kfs reactions were calculated using this method. Plots of K against P and T are given in Fig. 15.2. For both reactions, Powell's method results in lower pressures. These pressures may therefore be underestimates, but should provide a self-consistent framework in which to compare different areas within the Moines. In a number of rocks, particularly those affected by strong late deformation or retrogression, pressures obtained from the Na reaction are lower than the minimum pressure from the Kfs reaction. This may be due to underestimation of γ^{wm}_{pa}, or to loss of paragonite component from the white mica during low-grade recrystallisation.

Another equilibrium which can be used is:

$$\underset{\text{zoisite}}{4Ca_2Al_3Si_3O_{12}(OH)} + \underset{\text{qtz}}{SiO_2} = \underset{\text{plagioclase}}{5CaAl_2Si_2O_8} + \underset{\text{garnet}}{Ca_3Al_2Si_3O_{12}} + \underset{\text{fluid}}{2H_2O}$$

(e.g. Ghent et al. 1979)

For this reaction, using Helgeson et al.'s data,

$\Delta H^\circ_{1,298}$ = +56325 cal $\quad\Delta S^\circ_{298}$ = +103.738 e.u. $\quad\Delta V_s$ = +1.5048 cal/bar

$$K = \left(\frac{\left(a^{pl}_{an}\right)^5 \cdot \left(a^{ga}_{gr}\right)^3}{\left(a^{ep}_{zo}\right)^4 \cdot \left(a^{qtz}_{SiO_2}\right)}\right) = \left(\frac{\left(X^{pl}_{Ca}\right)^5 \cdot \left(X^{ga}_{Ca}\right)^3}{\left(X^{ep}_{Al}\right)^4}\right)\left(\frac{\left(\gamma^{pl}_{an}\right)^5 \cdot \left(\gamma^{ga}_{gr}\right)^3}{\left(\gamma^{ep}_{zo}\right)^4}\right)$$

No distinction was made between zoisite and clinozoisite, and ideal solution was assumed for this phase (cf. section 15.1.1). Curves were calculated using Powell's (1978) method, and the results were comparable with those obtained using other equilibria. The effect of low water pressure would be similar (i.e. $X^{fl}_{H_2O}$ = 0.5 gives a 1 kb increase in pressure).

15.1.4 Comparison with other geobarometers

The only published geobarometer which is applicable to Moine assemblages is that of Ghent & Stout (1981). They used the reactions:

$$\underset{\text{white mica}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{garnet}}{Mg_3Al_2Si_3O_{12}} + \underset{\text{garnet}}{Ca_3Al_2Si_3O_{12}}$$

$$= \underset{\text{plagioclase}}{3CaAl_2Si_2O_8} + \underset{\text{biotite}}{KMg_3AlSi_3O_{10}(OH)_2} \qquad (1b)$$

$$\underset{\text{white mica}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{garnet}}{Fe_3Al_2Si_3O_{12}} + \underset{\text{garnet}}{Ca_3Al_2Si_3O_{12}}$$

$$= \underset{\text{plagioclase}}{3CaAl_2Si_2O_8} + \underset{\text{biotite}}{KFe_3AlSi_3O_{10}(OH)_2} \qquad (2a)$$

These were calibrated for natural assemblages against the garnet-sillimanite-anorthite-quartz geobarometer, using Ferry & Spear's (1978) geothermometer. This yielded the equations:

$$0 = -8888.4 - 16.675T + 1.738P + RT\ln K_{Mg} \qquad (1b)$$

$$0 = 4124.4 - 22.061T + 1.802P + RT\ln K_{Fe} \qquad (2a)$$

where K_{Mg} and K_{Fe} were calculated assuming ideal solution.

To make this calibration compatible with the others used in this chapter, it is necessary first to calculate the pressure using Ferry & Spear's (op.cit.) thermometer, then correct it to Thompson's (1976) temperature along the slope of the garnet-sillimanite-anorthite-quartz barometer. Ghent et al. (1979) used the equation

$0 = -11676.5 + 32.819T - 1.30064P - RT\ln K$ for the latter.

This can be rearranged to give:

$$P = \frac{(32.819 - R\ln K)T}{1.30064} - \frac{11676.5}{1.30064}$$

or $P = mT - 8977.5$ where the slope m is a constant for a given value of K.

The original pressures from equations (1b) and (2a) are presented in Tables 15.4 - 15.7 as P (1b) and P (2a), and the corrected pressures as (P1b)* and (P2a)*. Agreement with the other barometers is good at 550-600°C, but poor at higher and lower temperatures. The higher pressures indicated for low grade Morar Division rocks are plausible, particularly since γ^{wm}_{pa} may have been underestimated, but the lower pressures at higher temperatures seem less likely. All likely errors would have the opposite effect, while most of the rocks would move well into the sillimanite field on Ghent and Stout's (1981) calibration.

Reactions (1b) and (2a) are useful, since they are relatively independent of the variation in composition of one mineral (the Na reaction is highly dependent on muscovite, and the Kfs reaction on plagioclase) and independent of $X^{fl}_{H_2O}$. In Fig. 15.3, they have been calibrated against the most reliable pressures from the Na, Kfs and zoisite reactions (neglecting rocks in which these give mutually contradictory results). The rocks used were: MAL49c, MAL37, MAL57, I1 (Kfs only), BK32, I30, K3, LH8b, LH13, LH15, R9, S6, Q43, M6. Plots of $(-RT\ln K - P\Delta V_s)$ vs. T give a reasonable straight-line correlation of the form: $(-RT\ln K - P\Delta V_s) = mT + c$. A least-squares regression, with T as the independent variable, gives the following results (standard errors in brackets):

(1b) $m = -49.019$ ($\pm$1.485) $c = +19525.37$ ($\pm$1268.22)

(2a) $m = -59.307$ ($\pm$1.672) $c = +36312.77$ ($\pm$1427.77)

These yield the equations:

(1b) $1.738P = T(49.019 - 1.987\ln K) - 19525.4$

(2a) $1.802P = T(59.307 - 1.987\ln K) - 36312.8$

The difference, (2a) - (1b), corresponds within errors to Thompson's garnet-biotite geothermometer (cf. section 15.1.2).

(2a) - (1b) $0.064P = T(10.228 - 1.989\ln K) - 16787.4$

Pressures from this calibration are tabulated as P(1b)‡ and P(2a)‡. They are distinctly lower at low temperatures, reflecting the steeper positive slope of this calibration, compared with Ghent & Stout's (op.cit). In rocks such as (59827), where the Na reaction gives anomalously low pressures, those from P(1b)‡ should be more reliable. However, since there is no correction for activity coefficients, the pressures are subject to an additional error, beyond the analytical error (probably up to 0.5 kb, by inspection of the variations between (1b)‡ and (2a)‡, and the reliable Na, Kfs and zoisite pressures - this is also reflected in the scatter in Fig. 15.3). Both these reactions have a strong positive slope (34 barK^{-1} for (1b)‡, 54 barK^{-1} for (2a)‡, at 600° and 5kb; decreasing at low pressures).

A number of possible reasons exist for the disagreement between Ghent & Stout's (1981) calibration (also shown on Fig. 15.3) and the present one. There are probably significant differences in activity coefficients between the two sets of rocks, since the values for γ quoted by Ghent & Stout (op.cit.) elsewhere in their paper are very different from those calculated in this chapter, despite the use of similar solution models. Alternatively, Ghent et al.'s (1979) garnet-sillimanite-anorthite-quartz calibration (which rests on the assumption that their kyanite/sillimanite isograd corresponds to Holdaway's (1971) experimental determination of the kyanite/sillimanite transition) may be in error. Note that since the P,T slopes of (1b)‡ and (2a)‡ are relatively close to those for the Na and Kfs reactions, the choice of geothermometer makes little difference to the new calibrations on Fig. 15.3.

15.2 P,T estimates

15.2.1 Introduction

P,T estimates were made using Thompson's (1976) garnet-biotite geothermometer, and the geobarometers discussed in sections 15.1.3 and 15.1.4. The primary calibration was made using the Na reaction calculated by Powell's (1978) method, with the Kfs and zoisite reactions where the appropriate minerals were present. In rocks where the Na reaction was affected by retrogressive changes in white mica composition, pressures calculated from equations (1b)‡ and (2a)‡ were used. These calibrations give the lowest pressures of all those available, so although different areas should retain the same relative position, all the pressures will have to be increased if one of the other calibrations proves to be a better choice.

Relative percentage errors were estimated from 2σ errors in the apparent concentrations for the appropriate elements in typical microprobe analyses. These will be minimum errors, since they do not take into account errors in the ZAF correction or variation in the detector response curve during the day. However, in most cases, the analytical error will be dominated by that in the apparent concentration. For the garnet-biotite thermometry, from inspection of Fig. 15.1, errors of about ±30°C result at 500-600°C. Errors for the geobarometers are quoted in Table 15.3. These are errors in the placement of the geobarometer curve, and not those propagated from errors in the temperature estimate. Realistic errors in pressure for a given estimate would be about double those in Table 15.3 (cf. the slopes of the geobarometers in Figs. 15.5-15.12), i.e. c. ±1kb. This is the precision of the pressure estimate; due to the wide range in pressures obtained from alternative calibrations, it is not possible to give a realistic value for its accuracy.

15.2.2 Morar Division

The main study in the Morar Division was based on the Mallaig area (Chapter 3), where the early metamorphic textures are well preserved. A number of other samples were taken from west of the Sgurr Beag Slide, for comparison with the adjacent Glenfinnan Division.

(i) Mallaig

Data from Mallaig are summarised in Table 15.4 and Fig. 15.5. Garnets are essentially MP1 and MS2 - MAL37(in) and (edge) are from a garnet showing classic Morar Division zoning (Fig. 15.4; cf. MacQueen & Powell, 1977), but MAL49c may have some slight MP2 growth, and MAL51(edge) from a calc-silicate is MP2. The results cluster very closely around 520°C at 6-7kb. MAL49c(c) and MAL57(c) give rather higher temperatures, possibly indicating higher grade during MP1, although they still lie within the 2σ range. MAL51(c) gives a similar pressure, and the inclusion-based temperature should be reliable. MAL37(c) gives an anomalously low pressure, but MAL51(e) is probably a good estimate for MP2, which was expected to be at lower temperatures. MAL49c(e) may relate to an intermediate stage of cooling (its reverse zoning in MnO is probably best explained by growth during declining temperatures - cf. Tracy et al. 1976). In summary, conditions passed from (?higher temperature) MP1, through MS2 at 520°C, 6.5kb, to MP2 at c. 460°C, 5.8kb.

(ii) Inverie

Only one rock (I1) was studied from Inverie (10km east of Mallaig, across the Morar Antiform; cf. Chapter 4). It is a K-feldspar-bearing semipelite carrying small, Ca-rich garnets, and is comparable to MAL57. Results are summarised in Table 15.4 and Fig. 15.6. The Na reaction gives a much lower pressure than the Kfs reaction; the latter agrees with reactions (1b)‡ and (2a)‡. Detailed textures cannot be determined, due to the presence of an intense S3 fabric, related to the Knoydart Slide. The muscovite has been thoroughly recrystallised by this low grade event, and its anomalously low Na_2O content ($<$0.3%, and below detection limit, Appendix 2) is responsible for the high value of K_{Na}. The most realistic P,T estimate is 513°C at 5.7kb.

(iii) Ardnamurchan

Three samples collected by B.H. O'Brien were studied (Table 15.4, Fig. 15.6). Their locations are shown on Fig. 15.17. (59828) and (59827) are garnetiferous pelites from either side of the en echelon continuation of the Morar Antiform, and (59564) is a garnet-poor semipelite from just west of the Salen Slide (Chapter 6). In this area, major late-tectonic slides disrupt the stratigraphy (O'Brien 1981), so the metamorphic pattern might be expected to be more complex than to the north, although Butler (1965) recorded a general west-to-east increase in grade.

All these rocks have suffered strong D3 and D4 deformation, so results using the Na reaction in particular should be treated with caution. (59828) edge suggests conditions of c. 470°C at 4-4.5kb. The core gives extremely low temperatures, probably due to loss of FeO from biotite as the garnet grew. For (59827), the Na reaction again gives low pressures. Results of 580°C at 6 - 6.5kb from (1b)‡ and (2a)‡ are distinctly higher than any found at Mallaig, and the core composition may indicate higher temperatures still. (59564) is higher again, but the 2σ ranges overlap considerably, although pressures may be somewhat lower. The mean P,T conditions for both rocks are about 600°C at 6kb. Clearly there is a very distinct difference between rocks east and west of the Morar Antiform, with the latter metamorphosed at a shallower level. This could favour a model involving thrusting although more samples would be required to rule out a continuous variation.

(iv) Kinlochourn

One rock was studied (LH23), from the semipelites west of the Sgurr Beag Slide. It contains small, round garnets, augened by the S2 slide fabric (cf. section 5.3.2). Garnet edge compositions give results which are comparable to the eastern sample from Ardnamurchan, i.e. c. 590°C at 5.5 kb (Table 15.4, Fig. 15.5). Next-to-edge compositions give c. 650°C at 6.5 - 7kb. This seems unrealistic in view of the relatively fine grain-size and lack of migmatisation in these rocks, in comparison with the nearby Glenfinnan Division. A similar pattern is seen in the Knoydart Pelite, i.e. next-to-edge garnets give very high temperatures and pressures; this may be significant, since, pre-sliding, the Knoydart Pelite probably lay a relatively short distance to the west of these rocks (cf. sections 14.2.3. 18.4). Such a pattern could result if some Fe-rich phase (e.g. ilmenite, staurolite, phengite) was consumed during growth of the outer zones of the garnet, and contributed Fe to garnet and/or biotite.

(v) Ben Klibreck

The semipelite studied here (BK3) was taken from some distance west of the main slide, to avoid subsidiary slides which occur in this area. In general, garnets are augened by the S2 fabric here, but the relations are ambiguous, due to strong recrystallisation of the matrix. This garnet is probably comparable to the early garnets further south, but it is possible that it is Caledonian (cf. section 9.5.3). Next-to-edge garnet gives results of c. 614°C at 7kb, somewhat higher than the LH23(edge) values (Table 15.4, Fig. 15.5). BK3(edge) gives very high temperatures and pressures. It is possible that this is a late overgrowth, related to the widespread Caledonian migmatisation and granite injection in this area, and that equilibrium was not attained.

(vi) Conclusions on Morar Division P,T conditions

The results obtained are summarised below, and shown on Fig. 15.13:

Mallaig (MP1 550°C?); MS2 520°C, 6.5kb declining to 460°C, 5.8kb in MP2.

Inverie 513°C, 5.7kb.

Ardnamurchan 470°C, 4-4.5kb (west); 585°C, 6-6.5kb (central); 610°C, 6kb (east).

Kinlochourn 590°C, 5.5kb.

Ben Klibreck 615°C, 7kb.

They clearly fall into two groups - one immediately west of the Sgurr Beag Slide, and a lower-grade set 10-20km further west. Unfortunately, the invervening area lacks suitable lithologies to test whether or not this reflects a continuous variation in metamorphic grade, as suggested by Kennedy (1949) and Tanner (1976).

15.2.3 Glenfinnan Division

The Glenfinnan Division rocks studied come mainly from just above the Sgurr Beag Slide or comparable structures, and will initially be described on a north-to-south basis. The Knoydart Pelite will be discussed with the Glenfinnan Division (cf. sections 4.4, 14.2.3). Several more easterly areas were sampled: Loch Quoich, the Strontian granite aureole and the "old Moines" of

Loch Ruthven, southeast of the Great Glen Fault. In these areas, later deformation and metamorphism is intense, so interpretation of the results is more difficult. The sample locations are shown on Fig. 15.17.

(i) Ben Klibreck

The rock chosen was a migmatitic pelite (BK32) from above the main slide. Garnets predate the main foliation, which is, however, coarsely recrystallised. They are probably broadly contemporaneous with the migmatisation. Edges of two separate garnets give results whose errors overlap strongly (Fig. 15.7, Table 15.5): the mean of these, 640°C at 6.4kb, is probably a reasonable estimate of conditions during migmatisation. A biotite inclusion near the core of one garnet gives about 600°C at 5.5kb; since the garnets are Ca-poor (Appendix 2), both pressure and temperature estimates may be reliable.

(ii) Sguman Coinntich

Two migmatitic pelites were studied from this area (K3 and K4; Fig. 15.8, Table 15.5). Both garnets and migmatitic *lits* are augened by a strong $S2_s$ foliation. The garnets exhibit little chemical zoning, but have a ring of coarse quartz inclusions near the edge; this may be broadly synchronous with migmatisation (section 8.3, 8.4). In K3, garnet 1 (core) was paired with biotite 1 included in a migmatitic plagioclase (Figs. 15.4, 15.8). Garnet 3 (in) represents the inclusion-rich zone, and was paired with biotite 2 at the edge of a *lit* plagioclase. Garnet 4 (edge) was paired with a coarse biotite in the pressure shadow. The rock contains many narrow zones of high late deformation, in which the biotite is fine-grained, Fe-rich and Si-poor (see Appendix 2); these were avoided. Matrix plagioclase from one of these areas is more calcic than that in the *lits*; this could be a primary feature, or may be related to the retrogression. P,T values for both compositions are plotted on Fig. 15.8.

The best estimate for the peak of metamorphism is about 600°C at 6.5 kb; the "edge" results may trace the cooling path of the rocks (to 570°C at 6 kb). K4 is much more strongly deformed than

K3, and garnets are retrogressed to biotite/epidote/ore masses. One unaltered garnet was analysed; plagioclase was available in augened lits, but micas were all thoroughly recrystallised in the new foliation. Using matrix biotite gives results of 670-700°C at 8-10kb (Fig. 15.8), much higher than those for K3. The biotite is very Fe-rich (Appendix 2), probably reflecting the amunt of Fe-rich garnet which has been resorbed into the matrix. Surviving garnet will retain its original composition (given the evidence for slight diffusion in garnet, even at peak-metamorphic temperatures in the Moines), and so be out of equilibrium with the matrix biotite.

(iii) Knoydart

Samples from the Knoydart pelite, above the Knoydart Slide (Chapter 4) show a wide range of calculated temperatures and pressures (Table 15.5, Fig. 15.9). I30 is a migmatitic pelite, with garnets which are probably MP1 and show similar zoning to the Morar Division garnets at Mallaig. The migmatisation in Knoydart is MS2-MP2, while the metamorphic peak at Mallaig was MP1. Possibly, during MP1, the Morar Division and the Knoydart Pelite were at similar grade, but the later MP2 metamorphism and migmatisation was restricted to more easterly areas. I22 is a coarse, muscovite-rich pelitic schist, unaffected by the D3 (slide) deformation, with MP2 garnets. I22(edge) gives results of c. 580°C at 7kb, using (1b)‡, (2a)‡ and Kfs reactions. A Ca- and Mg-rich analysis from near the margin of a garnet gives a very high pressure (655°C at 11kb). I21 is a calc-silicate with spongy garnets, surrounded by biotite-free haloes which clearly overprint the S2 fabric. Plagioclase is very strongly zoned, and is filled with anhedral clinozoisite grains which may be secondary. In Table 15.5, the K values labelled "plag2" are calculated using an analysed plagioclase. The other values labelled "An_{65}" were obtained by assuming that the climactic assemblage contained no clinozoisite; since the An_{30-40} plagioclase contains 30 - 50% clinozoisite, the original Ca content probably corresponded to about An_{65}. Three temperature estimates are given, one based on a rim composition, and two next-to-rim. The former gives a very high temperature; the bleached zone in the matrix around the garnets presumably indicates slow diffusion rates for Mg and Fe, so equilibrium may not have been maintained. Assuming An_{65} for the plagioclase composition makes little difference to the pressures for the Na reaction, but reduces those for the Kfs reaction by $>$1kb, making them compatible both with the Na curves and I22. I9 is a biotite-rich migmatite melanosome with MP2 garnets; it carries F3 crenulations, but is not strongly deformed. Garnet shows a marked outwards increase in CaO (Appendix 2). No plagioclase was present in the slide. A very rough estimate of pressure can be made by assuming a plagioclase composition of An_{20}, which would be typical for a migmatitic lit. Clearly, this depends on the lit being in equilibrium with its selvedge. A wide range of values is obtained, which, however, brackets those for I21 and I22. The zoisite pressures (Table 15.5; not shown on Fig. 15.9) for I21 correspond very closely to those obtained by using "plag 2" in reactions (1b)‡ and (2a)‡. This suggests that zoisite may have been part of an equilibrium assemblage, as claimed by Tanner (1976) - though perhaps not the climactic one. There is a large scatter in P,T estimates for this area, but the most realistic estimate for the MP2 metamorphism is c. 580°C at 6-7kb; the MP1 metamorphism may have taken place at around 515°C, 6.5kb.

(iv) Kinlochourn

Three samples were chosen from the lower part of the Glenfinnan Division pelite, in areas of low $D2_k$ and $D3_k$ deformation (Table 15.6, Fig. 15.7). In all cases, garnet is essentially pre-$D2_k$, but a narrow, post-$D2_k$ rim could be present (cf. Figs.15.4, 6.5(c)). LH8b is a migmatitic pelite with two types of garnet. One is subhedral, with rare inclusions, while the other is large, irregular and skeletal, with coarse quartz, biotite and plagioclase inclusions, and is probably syn- to post-migmatitic. Its composition corresponds to a position between the edge, and half-way to the core, of the subhedral garnet (Appendix 2). It is also quite uniform, over an area larger than the whole subhedral garnet, suggesting faster growth for the skeletal garnet. The skeletal garnet gives a P,T estimate of c. 650°C at 5.5-6kb using biotite and plagioclase inclusions. The garnet edge gives much lower results (555°C at 4.1kb). The core of a subhedral garnet gives 582°C, using an inclusion biotite, although the pressure should be treated with caution, since matrix plagioclase was used. LH15 contains large, inclusion-free garnets, which are unzoned, except for a narrow rim. Next-to-rim garnet and matrix biotite gives similar results to LH8b(skeletal). The rim is identical to LH8b(rim). LH13 contains garnets which, like those of Sguman Coinntich, have a ring of inclusions near their margins. It gives some rather puzzling results. Pressures are 1-1.5kb higher than the others, although edge temperatures are similar to some results from LH8b and LH15. LH13 has been left out of the final estimate. To summarise, the best estimate of conditions during the $MP1_k$ migmatisation is c. 640°C at 6kb. Some garnet grew at an earlier stage, at around 580°C, possibly at about 5kb. Narrow garnet rims grew at 550-560°C and 4-4.5kb; this may have been during post-metamorphic cooling, but another possibility is that it represents some later (Caledonian) metamorphic event (it was concluded in section 5.3.2 that the $MP3_k$ metamorphism probably reached about staurolite grade).

(v) Acharacle

Four rocks were studied, S6 and S18 from the Salen Slide zone, R9 and R10 from further east. Garnet is fairly uniform in composition, except for a narrow rim (Appendix 2). In R9, next-to-edge garnet gives a result of c.600°C at 5.5kb, while the Mg-poor rim gives a result of 530°C at 3kb (Table 15.6, Fig. 15.10). Na and Kfs reactions agree closely. Maximum pressure can be obtained by applying the zoisite equilibrium to R10, a garnet+hornblende+anorthite-bearing calc-silicate. "Core" and "edge" results (the latter a thin ? retrogressive rim) are compatible with the pressures obtained from R9. Garnet-biotite temperatures were calculated for S18, taken from the centre of a garnetiferous amphibolite body. These range from 620-636°C in a large skeletal garnet. S6 is a non-migmatitic, plagioclase-rich pelite from the slide zone. The rock is very highly deformed, with individual garnet and plagioclase crystals augened, so there is some doubt whether an equilibrium assemblage is preserved. In addition, the muscovite is unusually sodic, so its composition may have changed during sliding. Two temperature estimates based on inclusion biotite should be reliable; these give 669°C for the core, and 630°C for a position half-way to the edge. S6(in) uses next-to-edge garnet and matrix biotite, and gives similar temperatures; S6(edge) gives a slightly lower temperature. Pressures also agree fairly well, although S6(in) is incompatible with R10(c). The main metamorphism, then, with the bulk of garnet growth, probably took place at 630-640°C, 6-6.5kb; early stages of this metamorphism may have reached 670°C. As at Kinlochourn, garnet rims indicate lower-grade conditions, which could relate to cooling, or to later Caledonian metamorphism (R9 and R10 come from the eastern zone of Caledonian metamorphism and local migmatisation - cf. section 6.4.2).

(v) Mull

The Glenfinnan Division on Mull occurs as an outlier, and has suffered only low grade Caledonian metamorphism. M6 is a sample from the migmatitic pelite of Scoor Bay (section 7.3), and shows no obvious effects of Caledonian deformation or metamorphism. Large garnets are $MP2_r$ in age, i.e. syn- to post-migmatitic (Fig. 15.4). Next-to-edge garnets give c. 680°C at 8kb; edge compositions give 650°C at c. 7kb (Fig. 15.11, Table 15.7). Small garnets of indeterminate age give similar results to the edge of the large garnet. M1, a semipelitic schist carrying an $S4_r$ crenulation cleavage, gives somewhat lower temperatures. The thermometry is based on inclusion biotite, so temperatures should be quite reliable; possibly the garnets grew relatively early (e.g. $MP1_r$), and so do not record the metamorphic peak. The Na reaction gives distinctly lower pressures, probably due to loss of Na from muscovite during MS4 recrystallisation. H15 is a specimen from a kyanite-bearing pelite east of the Ross of Mull granite. Unfortunately, it lies well inside the thermal aureole, so matrix compositions must be very suspect. Plagioclase analyses as albite on the microprobe, but U-stage determinations indicate a composition of about An_{20}; this value was used in the calculations. Temperatures range around 600°C, lying between those from M1 and M6. Pressures are extremely high, using both the kyanite and Na reactions. The high pressures for the Na reaction are due to the very high paragonite content of the muscovite, which may be a thermal metamorphic effect. The Kfs reaction gives lower pressures but (1b)‡ and (2a)‡ give intermediate results. Since M6 contains abundant garnet, even a thin rim could change the matrix compositions, so only the "edge" result will be used. The average of all the estimates has been taken although they lie right at the 2σ limits: c. 600°C at 6.5kb is similar to other areas, but can only be considered a very rough estimate.

(vi) Loch Ruthven

In this area (near Foyers, southeast of the Great Glen), probable Glenfinnan Division basement to "young Moines" occurs (A.J. Highton, verb. comm., Piasecki 1979). One kyanite-bearing sample, collected by A.J. Highton, was studied. It has suffered considerable deformation and retrogression, with garnet altered to biotite/epidote/ore masses, and then to chlorite, and kyanite altered to shimmer aggregate, which develops into muscovite porphyroblasts. Temperatures of 640-650°C (edge and next-to-edge), obtained using an unaltered garnet and matrix biotite, seem reasonable (Fig. 15.12, Table 15.6). However, pressures are impossibly low (well into the sillimanite field) when the Na reaction is used; this is probably due to low Na in the muscovite (below detection limit in one grain). Use of the kyanite reaction gives pressures of 7-8kb, while (1b)‡ and (2a)‡ give c.5kb. The latter pressure still seems slightly low (cf. Fig. 15.13).

(vii) Loch Quoich

This area has suffered intense Caledonian reworking and high grade metamorphism. However, garnets and migmatitic lits both predate all observed fabrics, and so a cautious interpretation of garnet-matrix compositions may give some indication of the conditions which prevailed during migmatisation. Garnet textural relations are shown in Fig. 15.4. Temperatures based on garnet core/inclusion biotite, and garnet edge/matrix biotite, range from 566-590°C (Fig. 15.12, Table 15.6). Pressures also cluster closely, with the Kfs reaction lying within the error bars for the Na reaction. P,T estimates of c. 590°C at 5.5-6kb are broadly comparable with other areas, despite the reworking.

(viii) Strontian

Two specimens from the Strontian granite aureole were studied (Chapter 13). These contain abundant cordierite, apparently replacing garnet, and sillimanite-K-feldspar (±muscovite) assemblages. S15 contained no garnet. S17 contained what appeared to be garnet relics in one of the cordierites. The garnet is extremely rich in Mn (Appendix 2), and separate grains, in different parts of an apparent cordierite pseudomorph, have similar compositions. It may be a part of an equilibrium cordierite-garnet assemblage (if it was a relic from an early garnet-rich assemblage, it would be expected to be poor in Mn). Ashworth & Chinner (1978) have suggested that garnet and cordierite were both stable in the Strontian aureole. Garnet/matrix biotite compositions give 620°C (Table 15.7), but Thomson's (1976) garnet-cordierite geothermometer gives 775°C. If Fe/Mg ratios in garnet-cordierite and garnet-biotite pairs are fixed at a given P,T, then the Fe/Mg ratio of cordierite-biotite pairs must also be fixed. The garnet and biotite in S17b are fresh, but the cordierite is intensely pinitised ($> 2.5\% K_2O$). This can perhaps be corrected by making the ratio $\frac{(Fe/Mg)\ cord}{(Fe/Mg)\ biot}$ in S17b, the same as that in S15, which contains fresh cordierite. This gives $(Fe/Mg)cord = 1.084$; with garnet (1), $\ln K_D = 2.08$, and $T = 620°C$. Taking this as the best temperature estimate, the Na, Kfs, (1b)‡ and (2a)‡ reactions give broadly similar pressures, averaging about 2.9kb. This is reasonable for a granite aureole, but garnet might not be stable at such low pressures. Low water pressure would increase tne Na and Kfs pressures, but (1b)‡ and (2a)‡ are independent of H_2O. The sillimanite reaction gives 5kb, which agrees closely with Ashworth & Chinner's (1978) pressure estimates. However, their temperatures are nearer to 700°C, and the positive slopes of all these barometers would add more than 2kb in 80°C, so that given a higher temperature estimate, the Na, Kfs, (1b)‡ and (2a)‡ reactions are actually more compatible with their garnet-cordierite barometry. Yardley _et al_. (1980) also obtained anomalously high pressures from the sillimanite reaction, in cordierite-bearing pelites from Connemara. Yardley _et al_. (op. cit.) also found that highly manganiferous garnets gave anomalously low garnet-biotite temperatures - from inspection of their fig. 7 the garnet-biotite temperature discussed above should be corrected to 650 - 670°C.

(ix) Conclusions

Preferred P, T values for the Glenfinnan Division are summarised in Fig. 15.13. The main analytical uncertainty in the values of K_D for the geothermometer is MgO in garnet; for the geobarometer, Na_2O in muscovite (±10% and ±30% respectively). Since different temperature estimates use different garnet compositions, taking the mean of several should impove the precision. However, different pressure estimates often use the same muscovite, so the mean of three or four pressure estimates on the same rock will have much the same uncertainty as the individual estimates. For this reason, the barometer curves can still only be considered precise to ±½kb. The results all cluster fairly closely around 625°C at about 6kb, lying virtually on the kyanite/sillimanite field boundary.

Some support for the accuracy of the pressure estimates is given by their relationship to the kyanite/sillimanite inversion (Fig. 15.13). Mull and Loch Ruthven both have widespread kyanite, but no regional sillimanite; no kyanite was observed at Sguman Coinntich, but it has been recorded in the general area (Clifford 1958b, Fleuty 1961, Tobisch 1963), while Kinlochourn has staurolite, kyanite and sillimanite, apparently crystallising in that order (Tanner 1976, p. 115). In Ardnamurchan, Butler (1965) reported one occurence of kyanite and staurolite, east of the Salen Slide, but the I.G.S. 6-inch sheets show several localities at which sillimanite was observed. Dalziel & Brown (1965) described several occurrences further east, and Stoker (1980) has mapped a sillimanite-K-feldspar isograd (probably pre-Caledonian) in southern Ardgour, east of the Strontian Granite. In these areas, it would be reasonable to expect heating through kyanite grade to sillimanite grade, with kyanite surviving as relics at the higher temperatures. The Knoydart Pelite appears to be slightly lower grade than the other areas, insofar as migmatisation is restricted to the garnetiferous pelite, and it has andesine-hornblende-biotite calc-silicate assemblages, as opposed to hornblende-anorthite at Acharacle, hornblende-bytownite at Kinlochourn - cf. Tanner (1976). Poole (1966) recorded kyanite in nearby northeast Morar.

Widespread sillimanite was recorded in the area east from Ben Klibreck by Read (1931) - i.e. in the Glenfinnan Division - while Soper & Brown (1971) recorded staurolite and kyanite in the (higher pressure) Morar Division to the west.

The internal consistency of the pressure estimates is shown by the close agreement between the Kfs and Na reactions for the K-feldspar-bearing R9, and of both with the maximum pressure based on the zoisite reaction. Where a large discrepancy does occur, it is usually explicable in terms of later reworking or metamorphism, and is consistent with the petrology of the sample.

Two areal groupings are seen, which may be significant in terms of the regional Precambrian P, T conditions. The three locations at the main outcrop of the Sgurr Beag Slide (R, LH, BK) give virtually identical results, as do the two western outliers of Mull and Sguman Coinntich. The Knoydart Pelite could be linked with either the easternmost Morar Division, or the westernmost Glenfinnan Division, but Loch Ruthven (AJH) is quite distinctive. The Quoich sample comes from a high level within the Glenfinnan Division, so would be expected to give somewhat lower P,T estimates than those from just above the Sgurr Beag Slide. These features will be discussed further in section 15.3.2.

15.3 Discussion

15.3.1 Comparison with melting reactions and origin of migmatites

Fig. 15.14 shows a net of melting and dehydration reactions, based on Thompson & Algor (1977). The lower temperature set of curves is for the system KNASH (K_2O-Na_2O-Al_2O_3-SiO_2-H_2O), while those at higher temperatures are for the Na_2O-free system KASH. The dotted curves refer to the muscovite-free system. They were derived by Thompson & Algor using a combination of thermodynamic reasoning, experimental data, and petrological evidence, and will serve as a simplified model of a pelite. The effect of CaO, and of $X_{H_2O} \neq 1.0$ can be assessed qualitatively, while ferromagnesian phases should have little effect on the melting reactions at the relatively low temperatures encountered.

If $a_{ab}^{plag} < 1.0$, the melting curves involving albite will move to higher P and T, along the line (Alb) joining the invariant points (in this context, (Alb) means the curve belonging to the equilibrium which does not involve albite, i.e. in this case, Ms+Q+L = Kfs+Sill+V; written as Ms+Q+L (Alb) Kfs+Sill+V). This curve strictly only applies to the system KNASH (e.g. to albite solid solution in K-feldspar); to take account of the presence of anorthite in plagioclase, some additional curves must be generated (Fig. 15.15).

The system CKASH ($CaO-K_2O-Al_2O_3-SiO_2-H_2O$) has an invariant point analogous to that in KNASH; its position in Fig. 15.15 is that suggested by Tracy (1978). The curves surrounding this invariant point are based on Tracy (1978) fig. 11, with additional constraints from Winkler's (1976) melting curves, and from the assumption that they would broadly parallel the KNASH curves. The addition of CaO to KASH expands the L fields at the expense of An-bearing assemblages, as would be predicted from the solubility of CaO in the melt, but the effect is rather less than that of Na_2O. The Ms+Q+L field is larger in KNASH than in CKASH. This implies that, relative to CaO, Na_2O is more soluble in (Ms+Q+L) than in (Kfs+Sill+V); a conclusion broadly consistent with observations. In Appendix 2, Ca/K is broadly similar in white mica and K-feldspar, but Na/K is higher in white mica. The greater solubility of Na than Ca in L is attested to by the greater displacement of the (Ms) curves in KNASH.

The KNASH and CKASH invariant points are joined by the plagioclase-absent CKNASH reaction:

(K-Na-Ca)Ms+Q+L (Plag) (K-Na-Ca)Kfs+Sill+V.

This is the locus of all the invariant points in the system CKNASH (with Ca/Na specified in each case) where the assemblage Ms+Q+L+Plag+Kfs+Sill+V can exist in equilibrium. The plagioclase composition ranges from albite at the low-temperature end to anorthite at the high-temperature end. For a given plagioclase composition, a set of curves could be drawn, similar to those surrounding the KNASH and CKASH invariant points; as the plagioclase composition varied, the curves would slide up or down the (Plag) reaction. Such a set of curves would describe the instantaneous equilibrium of the system.

In a closed system, as dehydration or melting reactions proceeded, the plagioclase would become more calcic, so an increase in temperature would be required before the reaction could continue. These reactions would therefore take place across divariant bands, which would lie between the KNASH and CKASH reactions. This is indicated on Fig. 15.15 for a typical gneiss which originally had An_{30} plagioclase. Assuming that early melts had normative An_{10} plagioclase, and the composition of the last plagioclase was An_{50}, the bands might be expected to extend from one-tenth of the way across the field to half-way across. This is broadly consistent with Winkler's (1976) suggestion of a 20°C rise in the temperature required to produce significant amounts of melt, when an An_{30}-bearing gneiss is compared with an albite-bearing one.

Thus for a typical Glenfinnan Division pelite, melting temperatures would be about 20°C higher than those in KNASH (given that a reasonable amount of melt would have to be generated to produce a migmatite) and Ms+L assemblages would be restricted to somewhat higher pressures. To decide whether a P, T estimate lay in the melting field, it could be reduced by 20°C and 0.5kb, and then compared directly with the KNASH reactions.

If $X_{H_2O} < 1.0$, each set of curves moves along its respective (V) curve. In Fig. 15.16, an attempt has been made to show the effect on the An_{30} curves derived above, of $X_{H_2O} = 0.6$ (the value obtained by Ashworth & Chinner (1978) for the Strontian granite aureole). For clarity, only the centre of each divariant band has been shown. The new invariant point has been defined by the intersection of (V) with (L) for $X_{H_2O} = 0.6$ (data being lacking on the quantitive effect of $X_{H_2O} < 1.0$ on the melting reactions). Using the figures in Powell (1978, p. 135) as a guide, $X_{H_2O} = 0.6$ causes a 23°C reduction in (L) at 3kb, increasing to 31°C at 7kb. The resulting curves, shown in Fig. 15.16, have a large pressure error, due to the low angle of intersection. A small change in the slope of (L) would have a large effect on the position of the invariant point. However, it is clear that such a reduction in X_{H_2O} greatly increases all melting temperatures, and restricts the coexistence of muscovite and liquid to quite high pressures.

P,T estimates from Strontian (section 15.2.3(viii)) are shown on Fig. 15.16: that labelled (gar-biot, musc) was derived using the geobarometers developed in this chapter. Given the divariant nature of the muscovite dehydration reactions, and some error in P or T, this is in agreement with the petrological interpretation that the rocks are in the sillimanite-muscovite-K-feldspar zone (i.e. between the K-feldspar-in and muscovite-out isograds in sillimanite + plagioclase-bearing pelites), although the pressure may be somewhat low for garnet + cordierite coexistence. Since the barometer curves slope at 30-40°C/kb, a higher temperature would move the P,T estimates more-or-less parallel to the muscovite dehydration reactions. The (gar-biot, sill) estimate is clearly at much too low a temperature or too high a pressure.

Ashworth & Chinner's (1978) estimate, based on garnet-cordierite equilibria, lies at a higher temperature than the (V) reaction, so Kfs-free pelites should have partially melted regardless of X_{H_2O}, and any muscovite would have to be retrogressive unless X_{H_2O} was close to 1.0. No clear textural evidence for melting is preserved, and indeed Ashworth & Chinner only infer its likely occurrence from the P,T estimate. Note that melting with X_{H_2O} much less than 1.0 would depend on a suitable P,T path. Isobaric metamorphism (perhaps reasonable in a thermal aureole) would permit vapour-absent melting, but if muscovite dehydration had already occurred at lower pressure, then vapour-absent melting could not take place.

The question of whether the Glenfinnan Division regional migmatites are likely to be partial melts can be considered from two viewpoints - firstly, the mineral association, and secondly, the P,T estimates of section 15.2.3 in relation to the curves of Figs. 15.14-16.

A typical Glenfinnan Division pelite has the mineral assemblage qtz - plag (An_{20-30}) - musc - biot - gar - ilm. K-feldspar and epidote are extremely rare. Lits almost invariably consist of quartz and plagioclase, the latter indistinguishable in anorthite content from that in the host. Minor muscovite may be present, as well as late garnet and screens of biotite. The host has the mineralogy listed above, although in some highly migmatised rocks, quartz and/or plagioclase may be missing. Since most of the Glenfinnan Division rocks are either in the kyanite zone, or have passed through it into the sillimanite zone, a P,T trajectory of shallow positive slope, near to or just above the kyanite/sillimanite inversion, is likely (the ky-sill curve of Holdaway 1971 is shown on Figs. 15.5-16). In a K-feldspar-free rock, partial melting would not be expected to occur until the reaction boundary (Kfs) was crossed, at about 670°C for An_{30} plagioclase, at pressures from 5 to 10 kb. This would give rise to a granitic melt, i.e. one containing Kfs, and leave residual kyanite or sillimanite in the restite. Clearly this conflicts with the observed petrography of the rocks. Note also from Fig. 15.14 that the Alb + Q + V = L reaction lies at higher temperatures than the (Kfs) reaction, so Ms+Alb+Q would be unstable at temperatures where a molten tronhjemitic lit could exist (although Ms+Q, in the absence of albite, would be stable). Similar arguments would apply for anorthite-bearing plagioclase (Fig. 15.15).

Thus the petrographic conclusion would be that migmatisation must have taken place at temperatures below about 670°C, i.e. to the left of the (Kfs) curve. Tracy (1978) gave a more detailed treatment of reactions in this P,T region, but the conclusions for the limited range of lithologies encountered in the Moine would be unchanged.

The P,T estimates of section 15.2 agree with the above conclusion. All lie near or to the left of the (Kfs) curve in the Ca-free system (Fig. 15.14), using Thompson's (1976) geothermometer. Given the effect of anorthite solid solution in plagioclase (most migmatites having An_{20-30}) and of $X_{H_2O} < 1.0$, in moving the (Kfs) curve to higher temperatures, it is likely that the temperatures in all areas fell below those required for partial melting of K-feldspar -free pelites.

The Sguman Coinntich, Knoydart Pelite, Loch Quoich and Mull migmatites are all petrographically indistinguishable from the others, yet lie below even the "granite" solidus; it might be expected that a change in mechanism of migmatisation from subsolidus to partial melting would leave some textural evidence in the rocks.

The only way in which a molten trondhjemitic lit could exist in a quartz+muscovite+plagioclase-bearing host (at pressures below 10kb) would be if X_{H_2O} was higher in the lit than in the host. Even then, the trondhjemitic material would have to be introduced from outside, since the host could not melt to produce a Kfs-free liquid. Such a model is inconsistent with the evidence for a local, isochemical derivation for the migmatites (sections 16.2, 17.1).

Thus a sub-solidus origin is favoured for at least the trondhjemitic regional migmatites.

Several of the P,T estimates on Fig. 15.14 lie above the "granite" minimum melt curve (ky) or (sill). Thus, in principle, granitic melts could coexist with Kfs-free, muscovite + quartz + plagioclase-bearing palaeosomes. Some regional migmatites do contain a small amount of Kfs in their lits; However, in every case, Kfs also remains in the matrix, which would not be expected in the simple system of Fig. 15.14. It might occur if melting preferentially removed albite from the plagioclase, leaving that in the palaeosome more anorthitic, and raising the equilibrium temperature for the (ky) curve; however, no such differentiation of plagioclase compositions was observed in the normal regional migmatites. In some exceptional situations, this does appear to have occurred - e.g. Q22, a Loch Eil Division pelite which may have undergone Caledonian partial melting (sections 16.3.2, 17.2). Similar lithologies (below the present erosion level) might have undergone partial melting during the Precambrian metamorphism to produce some of the large, coarse-grained, K-feldspar, tourmaline and beryl-bearing pegmatites which cut the Glenfinnan Division.

15.3.2 Regional P,T distribution

The main points of interest are the contrasts between the Morar and Glenfinnan divisions, and variations within each division. All the estimates refer texturally to the early (Precambrian) metamorphism. For most areas, pressures are fairly constant at 6kb+, indicating that the higher temperatures recorded in the Glenfinnan Division reflect a real difference in geothermal gradient, rather than simply deeper burial.

The Glenfinnan Division resting on the main outcrop of the Sgurr Beag Slide shows very uniform P,T conditions - this is consistent with the slide being a thrust, bringing up rocks from a fairly constant depth. The Loch Quoich rocks at the eastern (structurally and stratigraphically higher - cf. Brown _et_ _al_. 1970) margin of the Glenfinnan Division, give lower P and T, which fits with the Precambrian isograds being initially sub-horizontal and sub-parallel to bedding, as in Ardgour (Stoker 1980). Thus the relatively low grade areas in the higher Loch Eil Division to the east need not be part of a later cover sequence (as claimed by Lambert _et_ _al_. 1979). It also makes a local partial melting or metasomatic origin unlikely for the Quoich gneiss, since unlike the Ardgour gneiss, it does not lie in the area of highest grade.

Mull and Sguman Coinntich both lie west of the main Glenfinnan Division outcrop, and have lower temperature but higher pressure - however, a thrust cutting through already-folded rocks can cut locally down-section (as the Sgurr Beag Slide probably does on Mull - section 7.5; cf. also Allmendinger 1981). The Knoydart Pelite gives broadly similar pressures to the nearby Morar and Glenfinnan divisions, and has probably been brought along rather than up by the Knoydart Slide (originally gently - dipping, now dipping southeast at 40°); this slide could be the refolded equivalent of the Sgurr Beag Slide. Alternatively, the Knoydart Slide may be parallel to, but structurally underlying the Sgurr Beag Slide, so that the Knoydart Pelite forms an intermediate nappe (section 14.2.3 - cf. the Morar Division west of Ben Klibreck, lying between the Meadie and Naver (= Sgurr Beag) slides of V.E. Moorhouse (verb. comm.) and Soper & Barber (1982)).

Morar Division rocks immediately west of the Sgurr Beag Slide all give very similar temperatures, but with variable pressures; Kinlochourn has a distinctly low pressure, which is also shown at Inverie, 15km to the west. There is a marked difference across the Morar Antiform, both in Morar and Ardnamurchan, in supposedly the same stratigraphic unit (the Morar Pelite), indicating either that isograds (P and T) did not parallel the stratigraphy, or that the structure is more complex, involving thrusting or other duplication (note that due to the <u>en echelon</u> nature of the Morar and Ardnamurchan antiforms, Inverie is along strike from the <u>western</u> limb of the Ardnamurchan anitform).

The Glenfinnan Division always gives higher temperatures than the adjacent Morar Division, but pressure relations vary - presumably due to the slide cutting across Precambrian or early Caledonian structures.

In summary, the Morar Division shows a NW - SE increase in temperature within a given area, but may consist of several slices. The Glenfinnan Division also shows a northwesterly decrease in temperature at a given level, and an upwards decrease through the sequence. The difference between the two divisions decreases to the southeast - either the Precambrian metamorphism become uniform in this direction, or the slide cuts the isograds at a much reduced angle.

The MP3 event in Acharacle and Kinlochourn could be due to waning stages of the Precambrian metamorphism, or to thin Caledonian overgrowths on garnet; its restriction to areas of relatively high Caledonian grade favours the latter explanation, but more specimens would be needed to confirm this. In either case, the very low pressure could cause the marginal reverse zoning which is observed in plagioclase from these areas. If it is Caledonian, then the lower Strontian pressure would be more reasonable (it is unlikely that the relatively late granite was intruded at a greater depth than the main Caledonian metamorphism only 10km to the west).

CHAPTER 16

Bulk Chemistry

The composition of Moine metasediments will be discussed with respect to their mineral assemblages, using both new analyses and others collected from the literature. They will then be divided into migmatites and non-migmatites within the Glenfinnan Division, to look for any compositional controls on migmatisation. Finally, the chemistry of certain atypical rocks (migmatitic and otherwise) will be discussed, with particular reference to their trace elements.

16.1 A,K,F,M diagrams

In the absence of modes and mineral compositions for most rocks, only composition diagrams have been used, i.e. there is no correction for phases which are not represented in AKFM space (apart from apatite and calcite). A,K,F,M are defined in Fig 16.5. The AFM diagram is effectively projected from the A apex of the tetrahedron, and is not equivalent to the Thompson AFM diagram, projected from muscovite.

Collected analyses from the Morar Division are plotted in Fig. 16.1. Butler's (1965) garnetiferous pelite is a homogeneous, garnet and plagioclase-rich pelite, with biotite the dominant mica (cf. Mallaig, section 3.2). His "pure pelites" are muscovite-rich and garnet-poor, and often interbedded with psammites and semipelites. The garnetiferous pelites form a homogeneous group with constant M/FM, and fairly high F + M, reflecting the dominance of biotite over muscovite. "Pure pelites" have higher A values, but the AKF diagram shows that this is due to their muscovite content, i.e. the rocks are not really aluminous. Other pelites (mainly from Sutherland), show a broad scatter, particularly to higher M/FM ratios; psammites have distinctly lower A values, and very high K values, reflecting their arkosic, K-feldspar-rich nature. The rocks as a whole tend to have M/FM<0.5, and most lie on the K side of the muscovite-biotite join, while a few of the "other pelites" lie on or below the muscovite-garnet join - indicating that kyanite or sillimanite will not be formed until muscovite + plagioclase + quartz, or (in Fe-rich rocks) muscovite + garnet + quartz breakdown occurs (cf. Tracy 1978). The three

A-rich samples (45, 46, 47) are chiastolite hornfelses from the Carn Chuinneag aureole, and two of the cluster to the left of the muscovite-biotite join (23, 57) are staurolite-bearing.

A similar pattern is shown by the Glenfinnan and Loch Eil divisions in Fig. 16.2. Most of the pelites are comparable to the garnetiferous pelite, while semipelites and psammites are more A-poor and K-rich; almost all rocks have $M/FM < 0.5$. The kyanite- or sillimanite-bearing pelites are more A-rich than the others. The very A-rich sample is a kyanite-staurolite-sillimanite-bearing schist (42) from the Coire nan Gall series of Kintail, which may represent Lewisian basement. The other A-rich analysis (36) is of a sillimanite-garnet-biotite schist from the Glen Loy diorite aureole. A sillimanite gneiss from Moidart (28) plots in the muscovite-biotite-garnet field, and the fibrolite + muscovite-bearing Loch Coire migmatites (55, 103, 104) lie near to the muscovite-biotite join. The other Loch Coire migmatites are similar to the Glenfinnan Division, but the K-feldspar-bearing ones are distinguished on the AKF plot. It would appear therefore, that for the Fe-, K-rich Moines, rocks on the A side of the muscovite-biotite join can have kyanite, sillimanite or staurolite with muscovite; the slight overlap could be due to varying mineral compositions, analytical error, and failure to allow for such factors as Na in white mica, Fe in ilmenite or Fe^{3+} in micas.

29 new analyses from the Glenfinnan Division are plotted on Fig. 16.3. All the metasediments have $M/FM < 0.5$ (the four others are igneous), and all but Q22 have very similar M/FM ratios. The Quoich granitic gneiss (Q11a) is quite distinct from all the metasediments, and the psammites (LH18, Q49a, LH11) are all K-rich. From the preceding discussion, it is unlikely that any of these would have an aluminous mineral before muscovite breakdown; I24, S6 and particularly K4 are possibilities, although the last-named is weathered, and hence has a spuriously high Al/K ratio. In fact, only S15 and S17 have sillimanite, in qtz-plag-cord-biot-sill-Kfs(-gar) assemblages from the Strontian inner aureole, where muscovite breakdown has clearly begun (cf. Tyler & Ashworth 1982).

The compositional control of kyanite, staurolite and sillimanite can be tested more precisely in 3-dimensional AKFM space. Only those rocks which plot above the muscovite-biotite-garnet plane should carry kyanite, staurolite or sillimanite below muscovite breakdown in quartz-bearing rocks. It is apparent from Fig. 16.5 that all the new analyses plot well below this plane, but a few of the collected analyses (Fig. 16.4) may rise above it (for clarity, only the northern Morar Division has been plotted; all the southern samples lie well below the plane). This was tested, using a computer program developed by M.S. Brotherton, which determines if a given point lies above or below a plane defined by three AKFM compositions. Out of all the analyses in Appendices 1 and 3, only (26), a very siliceous rock in which slight weathering or analytical error could have a large effect on the A value, lay above the muscovite-biotite-garnet plane chosen. This suggests that the test is too strict; perhaps some Fe_2O_3 should be added to A, or FeO subtracted from F, or other mineral compositions chosen for specific rocks. However, (45), (46) and (47) lie close to the plane in the Morar Division diagram, as do (42) and (36) from the Glenfinnan Division, while (28) and the three fibrolite-bearing migmatites plot somewhat lower.

It would appear, therefore, that Moine metasediments in general will show no additional mineral phases between the garnet zone and the sillimanite-K-feldspar zone. This makes estimation of metamorphic grade difficult over most of the Moine outcrop, and is one reason why an extensive programme of geothermometry and geobarometry was undertaken (Chapter 15).

16.2 Compositional controls on migmatisation

16.2.1 Introduction

In the Glenfinnan Division, migmatitic and non-migmatitic rocks are interbanded on scales ranging from a few decimetres to a few metres (e.g. Plate 17.1(a)). Where the regional migmatitic lits lie in a fabric at an angle to bedding, they terminate at the edge of a migmatitic bed, and start up again when the cleavage intersects the next migmatitic bed. It is extremely unlikely that physically injected material could show such features, and in any case it is clear in areas of low strain (particularly Knoydart and the Ross of Mull), that the migmatite/non-migmatite alternation is in fact relict bedding. In Plate 7.2(b), the S2 fabric in which the migmatitic lits lie transects isoclinal F1 folds of migmatite/non-migmatite banding - i.e. the banding predates both the migmatisation and an earlier fold episode. In Plate 7.2(d), a thicker non-migmatitic bed is folded by F2, with axial planar migmatitic S2, with lits terminating against the folded bed. In Plate 7.2(c), S2 cuts across migmatitic and non-migmatitic beds, with segregations developed in the former; the bed to the left of the lens cap is graded, with the more pelitic top carrying small quartzofeldspathic lenses which pass into the normal migmatitic fabric. Other beds show varying degrees of migmatisation. It is therefore clear that some beds were more prone to migmatisation than others, presumably under the same P,T conditions; the most likely control is composition. Deformation may also play a part, since in these examples, incompetent pelitic beds are more highly migmatised. Elsewhere, certain pelites are non-migmatitic, while semipelites or psammites are migmatitic, so it can only be a secondary control.

In this section, the aim is to determine what sort of rock is most easily migmatised, not to look for evidence of metasomatism. Physical injection has been ruled out, and any metasomatic process would presumably by directed along the clear channels for fluid movement - the S2 fabric in the Mull rocks - so should produce banding parallel or perpendicular to S2, rather than parallel to original bedding, even in complex pre-existent structures. The question of large-scale metasomatism within the Moines has already been investigated by Steveson (1968) and by Brown (1967), but the latter author compared the Morar Division with the Glenfinnan Division, so the differences found were essentially stratigraphic, rather than metamorphic. Steveson compared the Morar Pelite in the garnet and sillimanite zones; although he could not rule out slight metasomatism, he did show that the amount must be very small - equivalent to a few percent or less of modal plagioclase - and so not sufficient to convert a non-migmatitic rock into a migmatitic one. The conspicuous plagioclase of the high-grade Morar Pelite is due to its coarse grain size. The low-grade Morar Pelite contains abundant fine-grained plagioclase, which has often been underestimated in the mode (e.g. Butler's (1965) chemical analyses are incompatible with his modes - almost twice as much plagioclase as recorded would be needed to provide the analysed Na_2O). Steveson studied rocks from the same statigraphic unit, in different areas; the present study is based on rocks from different lithological units within the same area.

16.2.2 A,K,C,N diagrams

Broad trends of compositional variation in migmatites and non-migmatites can be observed in the ACN and AK(CN) diagrams of Fig. 16.6. These effectively separate the "pelitic" and "feldspathic" portions of the rock. As expected from section 16.1, psammites have a higher K-feldspar content than pelites, but a similar An/Ab ratio; the migmatitic examples show no distinctive features. Within the pelites, the first thing to note is that there is no real tendency for migmatites to be more K-rich than non-migmatites, nor do they show any difference in An/Ab ratio. Experimental and theoretical work (e.g. Thompson & Algor 1977, Tracy 1978, Winkler 1976 pp. 314-318) indicates that partial melting in pelitic rocks should first affect those which carry K-feldspar, then plagioclase-bearing rocks in order of increasing anorthite content; the lack of such a trend is therefore evidence against a partial melting origin. However, the migmatites as a group tend to have a higher proportion of feldspathic to pelitic components (K4 probably has a spuriously high A value, due to weathering), and plagioclase appears to be more important than K-feldspar. A highly pelitic rock such as M3, with little or no feldspar (or quartz, in this extreme example), would be incapable of forming quartzofeldspathic lits by any isochemical process. However, some of the other rocks, with perhaps 10% modal feldspar are not migmatised, so some further factor must operate. These conclusions essentially bear out the field observation that the most highly migmatised rock is the garnetiferous pelite lithology (section 16.1), which often has 30% plagioclase. The significance of these and other differences will be investigated further using statistical methods.

16.2.3 Statistical study

(i) Major elements

The new analyses, and those collected from the literature in which it was clear from the original description whether or not the rock was migmatised, were grouped in various ways to test for differences in bulk composition between migmatites and non-migmatites. Unfortunately, most previous studies have concentrated on the migmatites, so analyses of unmigmatised rocks (which probably account for more than 50% by volume of the Glenfinnan Division) are relatively scarce.

The results are summarised in Table 16.1, and the means and standard deviations for each group presented in Table 16.2. Histograms for "all migmatites" and "all non-migmatites" are given in Fig. 16.7.

From inspection of Table 16.2, it can be seen that the rocks are all typical pelites or semipelites, with the only noteworthy feature being a fairly high Na_2O/K_2O ratio (approaching 1 in some groups), presumably indicating a generally immature sedimentary starting material. There also appear to be systematic differences between migmatites and non-migmatites. Within each group, the migmatites have higher SiO_2, CaO, Na_2O and lower Al_2O_3; this results in a higher ($Na_2O/Na_2O+Al_2O_3$)ratio for the migmatites. The new analyses show higher (Na_2O/Na_2O+CaO) in the migmatites, but this is not borne out by the collected analyses.

The significance of these differences was investigated, initially using the F and t tests. These compare, respectively, the variances (σ^2) and the means of two sets of analyses; they assume that the distributions are normal, and the t test requires that the two sets have the same variance.

Most of the groups are too small for it to be apparent from a histogram if the distribution is normal. Group VI (the largest) has been plotted on Fig. 16.7. In order to remove the effect of dilution by quartz, data recalculated on an SiO_2-free basis were also considered. For the migmatites, most elements give near-normal distributions. The main exceptions are FeO and MgO, which are too "spread out", and SiO_2 and Al_2O_3, which are skewed to left and right respectively, and may also be bimodal. This reflects the dominance of pelites over psammites in the analyses, and a lack of intermediate (semipelitic) compositions. The first is a sampling effect, while the second follows from the characteristic pelite/psammite striping of the Glenfinnan Division. On an SiO_2-free basis, Al_2O_3, FeO and MgO become more nearly normal, but K_2O becomes skewed to the left, due to overcompensation of arkosic psammites, which give the spread to high weight %. (Na_2O/Na_2O+CaO) and (Fe_2O_3/Fe_2O_3+FeO) are normal, but ($Na_2O/Na_2O+Al_2O_3$) is skewed to the right, with a sharp cut-off just before the limiting value of c. 30% (the ratio found in a typical plagioclase). The non-migmatites show essentially similar features, but with more irregular histograms, due to the smaller sample size. SiO_2 and Al_2O_3 are more nearly normal, due to the presence of some very aluminous pelites, but CaO is skewed to the left. On an SiO_2-free basis, all the elements are plausibly normal, except Al_2O_3 and T Fe_2O_3, which are skewed to left and right respectively. (Na_2O/Na_2O+CaO) and (Fe_2O_3/Fe_2O_3+FeO) are probably normal, but ($Na_2O/Na_2O+Al_2O_3$) is again skewed to the right. Most elements, then, should come sufficiently close to a normal distribution for the F and t tests to be applied.

The F test uses the ratio $F = (\sigma_1/\sigma_2)^2$, where σ_1 is the larger standard deviation; the result is compared with the critical value of F for a particular confidence level, at the appropriate number of degrees of freedom (tables in Siegel 1956). It can be seen from Table 16.1 that the non-migmatites show larger variance in most elements, particularly SiO_2, Al_2O_3, CaO, Na_2O and K_2O, presumably representing quartz, mica and feldspar components - the implication being that the non-migmatites are lithologically more varied, but probably have the same mineral assemblage. The collected migmatites form a particularly homogeneous group (cf. also Table 16.2), and represent the typical "garnetiferous pelite" lithology (cf. the new analyses, which were specifically selected to give a wide spread of compositions). On an SiO_2-free basis, most of the differences in variance remain, but those related to feldspars (Na_2O, CaO, K_2O) are reduced - i.e. SiO_2 dilution explains the variance of those elements in the non-migmatites, better than it does in the migmatites. An interesting feature is the consistently low variance of the oxidation ratio in the migmatites - this was also noted by Brown (1967), who attributed it to homogenisation by a metasomatising fluid, although it may simply be a function of the original sedimentary composition. In the present study, separate migmatitic beds have similar oxidation ratios. Enhanced diffusion within a migmatitic bed might homogenise that bed, but there is no reason why different metre-thick beds should have the same ratio, unless it was inherited. Steveson (1968) noted some tendency towards local homogenisation (over a few metres) <u>within</u> individual migmatitic units (as compared to similar low-grade units). The few instances in which the migmatites have greater variance are probably spurious (e.g. T Fe_2O_3 in Group I is dominated by one highly weathered sample).

The t test compares the means of two sample distributions, using the parameter $t = \dfrac{X_1 - X_2}{S_c\sqrt{(1/n_1 + 1/n_2)}}$

where $S_c = \sqrt{\dfrac{(n_1-1)\sigma_1^2 + (n_2-1)\sigma_2^2}{n_1+n_2-2}}$. Critical values of t are tabulated in Siegel (1956). The t test is unreliable in cases where the variances of the two distributions are significantly different, i.e. where there is a positive F test, particularly if n_1 and n_2 are also very different. In Group I, Na_2O, Al_2O_3 and (Na_2O/Na_2O+CaO) all show differences in means, significant at close to the 95% confidence level, while (SiO_2 -free) Al_2O_3, Na_2O and ($Na_2O/Na_2O+Al_2O_3$) show differences significant at better then 95% - in effect, the reduction in variance produced by correction for SiO_2 dilution enables smaller differences in the means to be detected. When psammites are included (Group II), only Na_2O and ($Na_2O/Na_2O+Al_2O_3$) remain significant, since this increases the overall variance. In Group III, significant differences in means are shown by a number of elements, at up to 99% confidence. However, in each case, there is a significant difference in variance, so these results are unreliable. Similar comments apply to Group IV; in both cases, the small number and large variance of the non-migmatites make the t test unsuitable. Groups V and VI should give the most reliable results, since these use the largest number of samples. They show significantly higher SiO_2 and Na_2O, and lower Al_2O_3 in the migmatites, at better than 99% in some cases. Migmatites also have higher ($Na_2O/Na_2O+Al_2O_3$) and (SiO_2-free) Na_2O, and lower (SiO_2-free) Al_2O_3. In addition, Group V migmatites have higher (SiO_2-free) FeO, and lower oxidation ratio, while Group VI migmatites have lower Fe_2O_3 and K_2O, and higher (SiO_2-free) CaO. Most of these results are, however, cast into doubt by the positive F tests.

An alternative to the t test is the Mann-Whitney U test (Siegel 1956 pp. 116-127). This a non-parametric test which does not require either that the distributions are normal, or that the variances are the same. It is less sensitive than the t test (i.e. it is more likely to reject a positive result), but for $n>20$, it is about 95% as powerful. The tests were carried out using a library program on UMRCC; the results are presented in Table 16.1 as the percentage probability that the two means tested are <u>the same</u>, i.e. the smaller the number, the more significant is the difference in the means. For Group I, the differences in Al_2O_3 are rejected, and that in Na_2O only remains significant for the SiO_2-free values (at 97.3%). ($Na_2O/Na_2O+Al_2O_3$) is also highly significant, while (Na_2O/Na_2O+CaO) is barely significant. For the collected analyses, Groups III and IV, all the significant t tests are rejected by the U test - particularly striking are Al_2O_3, which drops from 99% significance to 55%, and ($Na_2O/Na_2O+Al_2O_3$), which drops from 95% to 6%. This need not imply that the migmatites and non-migmatites are the same - it merely means that a much larger sample of non-migmatites would be needed to place any differences on a sound statistical basis. For Groups V and VI some more positive results emerge from the U test. Na_2O is highly significant, both "normal" and "SiO_2-free". Al_2O_3 comes close to 95% for the SiO_2-free data, but both it and SiO_2 only reach about 85% for the "normal" data. This still means, however, that there is only a 1 in 6 chance that the difference is due to random variation. ($Na_2O/Na_2O+Al_2O_3$) shows a more significant difference than does either element alone, and (SiO_2-free) CaO approaches 80% significance in Group VI. The higher FeO in the migmatitic pelites (significant at 93%) probably reflects a genuine lithological difference (abundance of garnet and biotite), although it must be incidental to the process of migmatisation.

In summary, the main differences are that the migmatites have more Na_2O and the non-migmatites more Al_2O_3; the fact that ($Na_2O/Na_2O+Al_2O_3$) is more significant than either element implies that the two are not behaving independently. In both cases, correction for quartz dilution reduces the variance, and increases the significance of the difference in means. Secondary differences appear to be higher CaO, FeO and SiO_2 in the migmatites, and possibly higher K_2O in the non-migmatites. The fact that the group with the most samples gives the most significant results, lends some further support to their reality. It implies that, in this largest group, the expression for t (and the analogous one for U) are dominated by n, and thus that further samples are not significantly increasing the variance - i.e. that the sample distributions represent well the respective populations. This is exemplified by the case of CaO. Since there is no difference in (Na_2O/Na_2O+CaO) ratio, it would be predicted that, if Na_2O is larger, CaO will also be larger. However, only in the largest group was this difference detected - in the others, the number of samples was not sufficient to distinguish a small difference in means, between two distributions with fairly large variances. The variance of CaO is larger, relative to its absolute value, than that of Na_2O, presumably due to larger relative analytical error, or participation in more phases (e.g. plagioclase, apatite, garnet, calcite).

These conclusions may appear only to state the obvious - viz. that, other things being equal, migmatites are more plagioclase-rich, and non-migmatites more aluminous (in practice, muscovite-rich). Although they do not point directly towards a mechanism for the migmatisation, they can be used as negative evidence. As with the AK(CN) diagrams (section 16.2.1), the absence of any correlation of (Na_2O/Na_2O+CaO) with migmatisation, (although Na_2O on its own is correlated) and the slight negative correlation of K_2O with migmatisation, provide powerful arguments against partial melting as a mechanism. The type of chemical contrast found within this generally migmatitic area is similar to that which has been used to invoke metasomatism (addition of Na_2O) when rocks from a migmatitic area have been compared with others from a non-migmatitic area (e.g. Brown 1967 in the Moines). It is therefore vital in all such studies that the high grade rocks are sampled randomly, or at least that samples are not restricted to migmatitic rocks. All Brown's (1967) samples were migmatitic, but there are many non-migmatitic semipelites on Ben Klibreck (cf. Chapter 9). It can be predicted that inclusion of these rocks would give a more Al-rich, Na-poor mean for the high grade area, which might then be comparable to the low grade area. The restriction of migmatisation, not involving partial melting, to particular bulk compositions also appears to be a feature of parts of the Dalradian, e.g. the Duchray Hill gneiss (Bradbury 1979), which cuts across metamorphic isograds and parallels the stratigraphy.

One possible application of these data would be to test anomalously fine-grained schistose pelites within the Moines. Most Loch Eil Division pelites are relatively fine-grained and non-migmatitic; this is one factor which has been used to infer that they are part of a low grade cover sequence to the Glenfinnan Division (cf. Lambert et al. 1979). If they fell into the non-migmatite group, this argument would be weakened.

Finally, the large disparity between some U and t tests emphasises the importance of bearing in mind the assumptions inherent in most statistical methods - in this study, it appears that differences in variance, as expressed in the F test, are more important than deviations from normality in ruling out the use of the t test.

(ii) Trace elements

Trace elements were determined in the newly analysed rocks, and are presented in Appendix 3. Means and standard deviations, grouped into migmatites and non-migmatites, are given in Table 16.3, with the results of F and t tests for Group I (pelites only) and Group II (all rocks) - cf. also Table 16.1 None of the trace elements give a positive t test, i.e. there is no significant difference between the means for migmatites and non-migmatites. The Mann-Whitney U test was therefore not applied, since its main function is to reject differences which are accepted by the t test. However, a number of elements do show a difference in variance.

Ba and Rb have greater variance in the non-migmatites - these elements are particularly concentrated in micas, so this feature reflects the large variation in mica content within the non-migmatites, contrasting with the relatively restricted composition of the migmatites. Ti shows greater variance in the non-migmatitic pelites, reflecting the variation in biotite and perhaps ilmenite content of these rocks.

Ce, Nd, Zr, Y and (almost) La, have significantly greater variance in the migmatites, despite the generally lower variance exhibited by the major elements. In sediments, these elements will be associated with detrital heavy minerals such as zircon, garnet, hornblende, allanite and monazite. The migmatites have immature, greywacke-like compositions (especially when compared with the micaceous non-migmatitic pelites), and might be expected to show a large variation in heavy mineral content, and also a higher absolute value, although the t test does not confirm this as significant. The migmatites are particularly dominated by the high values for Q22, which probably contains an exceptionally high proportion of heavy minerals (cf. section 16.3.2). If this sample was removed, the differences would no longer be significant.

Some ratios also give significant F tests. Ba/Rb has significantly greater variance in the migmatites when psammites are included, implying that feldspar is more important than mica; in particular, the relatively high K-feldspar content of Q49a and LH11 results in high Ba contents in these rocks. Rb/Sr shows much greater variance in the non-migmatites. Since Rb is mainly controlled by mica, and to a lesser extent by K-feldspar, and Sr by plagioclase, and to a lesser extent by K-feldspar, this reflects the large range of mica : feldspar ratios in these rocks (cf. Fig. 16.8). Ca/Y has a larger variance in the migmatites, due to the variation in Y content discussed above.

All the trace element variations are therefore explicable in terms of the mineralogy of the rocks themselves. As might be expected, the plagioclase-rich migmatites have a higher mean for Sr, while the micaceous non-migmatites have higher means for Ba and Rb.

However, due to the large variances, a positive t test was not obtained - a larger sample might have been able to confirm these differences. The trace elements do not help to distinguish between partial melting and subsolidus metamorphic differentiation, since the bulk composition of the rocks should remain unchanged, as long as melt did not leave the system. However, they do tend to argue against major metasomatism, since this would probably cause changes in the concentrations of mobile elements such as Ba and Rb.

16.2.4 Migmatite composition diagrams

The ACN and AK(CN) diagrams are quite effective in separating migmatites and non-migmatites, but if would be useful to have a diagram which related more closely to the modal composition of the rocks. Clearly, from the preceding section, any diagram should separate Na_2O and Al_2O_3; since plagioclase and muscovite are the most sodic and most aluminous minerals found in the majority of Moine rocks, these would form useful end-members. Quartz was chosen as the third apex of a triangular diagram, which restricts the composition to the system $(CaO-)Na_2O-K_2O-Al_2O_3-SiO_2-H_2O$, and involves making the reasonable assumption that the nature and abundance of ferromagnesian phases are relatively unimportant in controlling migmatisation. The pelites will thus be modelled as quartz-muscovite-plagioclase mixtures; corrections can be incorporated for phases which do not plot in this system. Because of the low molecular weight of quartz, all the data would concentrate at the quartz apex; this may be prevented by (say) dividing the SiO_2 value by 5, but a better solution is to recalculate the corners of the triangle to weight percent. The procedure followed is outlined below. Muscovite was taken to be pure $KAl_3Si_3O_{10}(OH)_2$ and biotite pure $K(FeMg)_3AlSi_3O_{10}(OH)_2$. The 'muscovite' apex was defined by $(Al_2O_3-CaO^*-Na_2O-K_2O)$, i.e. subtracting the Al_2O_3 in plagioclase, K-feldspar and biotite (CaO^* is corrected for calcite and apatite); the 'albite' apex was defined by (Na_2O), and the 'plagioclase' apex by (Na_2O+CaO). A version was also run with 'feldspars' represented by $(Na_2O+CaO^*+K_2O^*)$, but this gave a poorer separation of migmatites from non-migmatites. $K_2O^* = (K_2O-(muscovite)/3-(FeMg)O/6)$, i.e. it is corrected for muscovite and biotite and represents K-feldspar. The 'quartz' apex was defined by $(SiO_2-2CaO^*-6Na_2O-6K_2O)$. This allows for plagioclase, K-feldspar, muscovite and biotite. No simple correction can be incorporated for garnet; its FeO and MgO was treated as biotite, and the effect of significant garnet would be to inflate the 'muscovite' and 'quartz' values. The 'molecular' apices derived above were multiplied by the appropriate formula weights to give results in weight percent of the model minerals.

Fig. 16.8 shows the migmatites and non-migmatites of Group VI plotted using this method. Pelites plot towards the top of the triangles, psammites towards the base. The 'albite' and 'plagioclase' versions give similar degrees of separation. The migmatitic pelites form a compact group with slightly more plagioclase than quartz, while the migmatitic psammites cover a range of broadly 'arkosic' compositions. There appear to be two distinct trends. One is characterised by fairly constant quartz content, with variable feldspar/mica ratio, possibly reflecting various degrees of weathering of the mineral fragments other than quartz. The other is a straightforward quartz-plagioclase mixture, with low pelitic content, probably reflecting a variable source for very immature arkosic sediments. Some of the non-migmatites represent more mature (muscovite- and quartz-rich) sedimentary compositions, but others are indistinguishable from the migmatites. If lines were drawn parallel to one running from musc50/plag50 to qtz60/plag40, then in crossing them from top right to bottom left, there would be a progressive decrease in the number of non-migmatites, and an increase in the number of migmatites. There is no sharp cut-off between migmatites and non-migmatites - rather one can say that, for a given composition, there is a 50% chance that the rock will be migmatised, for another composition 90%, for another 10% and so on. To actually draw contours on the diagram would require many more samples of non-migmatites - something like 100 samples, collected at random from an area where migmatites and non-migmatites are well interbanded.

Fig. 16.9 shows some of Steveson's (1968) analyses. The samples represented by "all rocks" were collected at random, and should be numerically representative of the area in question (near Loch Eilt, in the Morar Division). The rocks consist mainly of fairly aluminous pelites and quartzose psammites; from personal experience of this area, most of the rocks are non-migmatitic.

Steveson also specifically collected migmatites from this area for a separate study - these are on the whole more albite- or plagioclase-rich, despite the fact that any large quartzofeldspathic veins were removed from the samples. The relative scarcity of such compositions in the randomly-sampled group reflects the relative scarcity of individual migmatitic beds, and emphasises again how severely biased would be any comparison of these migmatites with low grade rocks. The garnetiferous pelite corresponds very closely to the field of migmatitic pelites, which bears out the field observation (cf. Chapter 4) that this lithology is always the most highly migmatised. The larger number of semipelitic samples makes the migmatite boundary clearer, but there is still considerable overlap; however, a line running from musc60/plag40 to qtz75/plag25 would form an approximate limit to the migmatites.

In passing, it can be mentioned that Figs. 16.8 and 16.9 show the differences between the Morar and Glenfinnan divisions quite clearly. In the Morar Division, the rocks fall readily into the three groups of Butler (1965) - psammites (often quartzose), garnetiferous pelites and "pure pelites" (the plagioclase-poor group). In the Glenfinnan Division, the "pure pelites" are much rarer, and typical lithologies are the garnetiferous pelite and more feldspathic psammites. A distinctive muscovite-rich rock also occurs - cf. M3 from Mull, which may be a thin pelagic layer between turbidites (P.J. Brenchley, verb. comm.). This contrast explains why, in several areas of the Morar Division, just west of the Sgurr Beag Slide, the rocks are unmigmatised, even though temperature estimates are higher than those from some Glenfinnan Divison migmatites (cf. section 15.2.2).

16.2.5 Conclusions

Comparison of interbanded migmatites and non-migmatites shows that the main factor which makes a rock prone to migmatisation is a high Na_2O content, reflected in high modal plagioclase. High SiO_2 and CaO appear to be secondary factors, and low Al_2O_3 is also important, although this may be due to dilution of the pelitic component. High FeO and low K_2O also show a positive correlation, but these are probably incidental to the process of migmatisation (i.e. they are tied to high Na_2O for sedimentological reasons). There is no correlation with the presence of K-feldspar or albitic plagioclase, indicating that a partial melting origin is unlikely. Most of the Glenfinnan Division migmatitic pelites form a very homogeneous group, equivalent chemically to the garnetiferous pelite of the Morar Division. The other Morar Division pelites have compositions which make them unlikely to be migmatised, even at high grade.

There is a large overlap between the migmatites and non-migmatites, with the latter having a particularly wide spread. A number of possible explanations exist for the gradational nature of this boundary. Analytical error, and errors in the corrections on the diagrams, would cause random scatter. Failure to correct for garnet would move the garnetiferous pelite away from the plagioclase apex, towards the non-migmatites. In Fig. 16.8, variations in P, T and X_{H_2O} between the various areas may result in different cut-off points for the migmatites; the better separation shown in Fig. 16.9, based on one small area, supports this view. However, some of the non-migmatites from local areas (Kinlochourn, Strontian) are more plagioclase-rich than associated migmatites. It may be that some secondary effect (such as deformation) is required to trigger migmatisation; in that case, there should be a fairly sharp cut-off to the migmatites, but the non-migmatites could show a broader spread. There is some indication that this is so. If migmatisation took place by partial melting, one would probably expect it to be fairly rapid (cf. van der Molen & Paterson 1979; Wyllie 1977, and references therein), and for a given P, T and X_{H_2O} there would be a sharp cut-off. However, if sub-solidus diffusion was responsible, it would be a rate-controlled process, which might require geologically-significant time. Then, minor factors such as original grain size, nature and shape of grain boundaries, size and distribution of other minerals such as micas and graphite, influence of deformation or patterns of local fluid flow, could all effect the rate of growth of segregations, once the initial compositional requirements were satisfied.

16.3 Detailed discussion of selected rocks

16.3.1 Introduction

The chemistry of certain rocks, considered to represent phenomena distinct from the normal regional migmatisation discussed above, was investigated in further detail. The samples chosen included a possible partial-melt migmatite (Q22/Q22a); the West Highland granite gneiss (Johnstone 1975 - Q1, Q11a, Q13b); late, cross-cutting pegmatites (LH23/LH23a, Q31a); an intrusive felsic porphyrite (Smith 1979-Q42) and some gneisses whose meta-igneous or metasedimentary origin was not apparent in the field (Q49a, S12, LH11).

The norms for these rocks are listed in Table 16.4. The mesonorms were calculated by placing all corundum into muscovite, $KAl_3Si_3O_{10}(OH)_2$, and subtracting the corresponding amount from K-feldspar. Hypersthene was replaced by biotite, $K(FeMg)_3AlSi_3O_{10}(OH)_2$, with K-feldspar reduced and quartz increased accordingly. Zircon was calculated by assuming all Zr to be in $ZrSiO_4$. This should give more realistic Q, Or, Ab, An ratios for comparison with melting experiments in the granite system. Most of the high grade white micas are close to muscovite (Appendix 2); the presence of garnet, or octahedral aluminium in biotite, will result in over-correction of K-feldspar and quartz. The true values should lie between those for the CIPW norm and the mesonorm, but close to the mesonorm. Fig. 16.10 is a tetrahedral diagram showing Q, Or, Ab, An ratios in the CIPW norm; Fig. 16.11 shows triangular diagrams of Q, Or, Ab and Or, Ab, An ratios. The corresponding diagrams for the mesonorms are given in Figs. 16.12 and 16.13.

Fig. 16.14 shows melting reactions in the system K_2O-Na_2O-CaO-Al_2O_3-SiO_2-H_2O, with the divariant bands for An_{30} plagioclase shaded (for further explanation, see section 15.3.1). To this diagram have been added melting temperatures at 5 and 7kb water pressure, for rocks whose compositions lie close to the cotectic planes in the system Q-Or-Ab-An-H_2O. These were obtained from Winkler's (1976) figs. 18.8 and 18.9, with a reduction of 5°C to allow for muscovite saturation in all these melts, and a further 5-10°C to allow for the discrepancy between Winkler's melting temperatures, and those of Thompson & Algor (1977), on which Fig. 16.14 is based.

Table 16.5 is an attempt to gauge the distribution of rare earths between the constituent minerals of selected rocks. The values refer to the percentage of the total weight fraction of that element which resides in each mineral, not to its relative concentration in that mineral - e.g. 92% of the Eu in LH23 should be contained in feldspar, 4.9% in apatite, etc., Mineral/mineral distribution coefficients were estimated from the ratios of the mineral/melt distribution coefficients quoted in Hanson (1978), and mineral proportions from the mesonorms in Table 16.4. Likely errors in this procedure arise from variations in distribution coefficients with P and T, the possibility that some of the sub-solidus assemblages never reached equilibrium, and that some rare earths are contained in minerals such as allanite or monazite which cannot be determined from the norm. However, they provide some constraints on models which involve retention of some rare earths in inert, refractory minerals during melting or metamorphic segregation.

16.3.2 K-feldspar - bearing migmatitic pelite

(i) Major elements

Q22 is a migmatitic pelite from just west of the granitic gneiss, in the spillway at Loch Quoich (Figs. 11.1, 11.2). It is typical of the Loch Eil Division pelites which lie on both sides of the granitic gneiss in this area, but is unusual in that its migmatitic lits are K-feldspar-bearing. Its petrography will be described more fully in section 17.2, but briefly, it is biotite-rich, garnet-free rock containing thick (several centimetres) quartzofeldspathic veins of 1-5mm grain-size. The feldspar is coarsely antiperthitic (c. 25% K-feldspar by volume), with composition about An_{25}. The host pelite contains numerous lenticles (a few millimetres by a few centimetres) of similar composition, in a biotite-muscovite-quartz-plagioclase (non-perthitic, An_{30}) matrix. From Figs. 16.12, 16.13, the vein Q22a lies close to a cotectic composition, in the plagioclase field, and the host, Q22, rather further away in the quartz field. From section 17.2, it seems likely that the vein, whether partial melt or metamorphic segregation, consisted essentially of quartz, plagioclase and K-feldspar, and restite consisted of quartz, biotite, muscovite and plagioclase. The small segregations in Q22 represent entrapped vein material, and the biotite-rich schlieren in Q22a represent restite, probably broken off the wall of the vein.

The moderate fractionation in plagioclase composition between Q22 and Q22a (An_{32} vs. An_{26}) and restriction of K-feldspar to segregation material tend to favour a partial melting origin. Melting of a rock similar to Q22 would begin on the Plag+Q+Or+L+V cotectic line; when all the K-feldspar was used up, the melt composition would move along the Plag+Q+L+V cotectic plane until it reached the projection from Q of the original bulk composition. Since plagioclase in Q22a is much too calcic to lie on the cotectic line, the melt composition must have moved into the plane, so all K-feldspar component should be contained in the melt. Temperatures for complete melting of the felsic components in Q22 and Q22a are given in Fig. 16.14. Below 7kb, the muscovite in Q22 and other plagioclase-bearing pelites would undergo dehydration melting at Q22a melting temperatures, which would be inconsistent with the petrography.

Certain minerals and elements, which would be expected to remain in the restite, have similar ratios between Q22 and Q22a (cf. biotite, muscovite, magnetite, ilmenite, apatite, zircon, Y - and rare earths which will be discussed further below). If it is assumed that Q22 and Q22a are both mixtures of melt and restite, one restite-rich, the other melt-rich, the proportions of melt and restite in each can be calculated. K-feldspar was assumed to be entirely contained in the melt, and apatite (taken as typical of the restite minerals) to be entirely contained in the restite. The model used is shown diagrammatically in Fig. 16.12.(b). The mass balance calculation is set out below.

Host		Vein	
$W^h_{ap} = W^r_{ap} . W^h_r$	(1a)	$W^v_{ap} = W^r_{ap} . W^v_r$	(1b)
$W^h_{Kf} = W^m_{Kf} . W^h_m$	(2a)	$W^v_{Kf} = W^m_{Kf} . W^v_m$	(2b)

i.e. $0.0070 = W^r_{ap} . W^h_r$ (1a) $\quad$ $0.0016 = W^r_{ap} . W^v_r$ (1b)

$0.0472 = W^m_{Kf} . W^h_m$ (2a) $\quad$ $0.1331 = W^m_{Kf} . W^v_m$ (2b)

Dividing (1a) by (2a) $\dfrac{0.0070}{0.0016} = \dfrac{W^h_r}{W^v_r}$ i.e. $4.375 W^v_r = W^h_r$ (3)

Dividing (1b) by (2b) $\dfrac{0.0472}{0.1331} = \dfrac{W_m^h}{W_m^v} = \dfrac{1-W_r^h}{1-W_r^v}$

i.e. $0.35462(1-W_r^h) = (1-W_r^h) \therefore 0.35462 - 0.35462W_r^v = 1-W_r^h$

and $-0.64538 - 0.35462W_r^v = -W_r^h$ (4)

Adding (3) and (4) $-0.64538 + 4.020W_r^v = 0 \therefore W_r^v = 0.1605$

$$W_m^v = 1-W_r^v = 0.8395$$

Substituting into (3) $W_r^h = 0.7022$, $W_m^h = 1-W_r^h = 0.2977$

i.e. Q22 is 70.2% restite, 29.8% melt

Q22a is 16% restite, 84% melt.

The composition of Q22 is particularly interesting, since it is at about this proportion of melt that the strength of a crystal mush changes by several orders of magnitude (van der Molen & Paterson 1979). Thus if the bed produced more than c. 30% melt, the crystal mush would be very weak, and excess melt could be squeezed out to form discrete veins, until the strength of the rock matched the imposed stress.

In the outcrop, thick veins make up about 1/6 of the total rock. If this ratio is accepted, then the relative proportions of restite and melt in the bed as a whole can be calculated:

Restite forms $\dfrac{5W_r^h + W_r^v}{6} = 61\%$ of the total

Melt forms $\dfrac{5W_m^h + W_m^v}{6} = 39\%$ of the total

The bulk compositions of host and restite can now be calculated; Q in the mesonorm will be given as an example. The results are listed in Table 16.6 (Q22r = restite, Q22m = melt).

$$W_Q^h = 0.3130 \qquad\qquad W_Q^v = 0.3117$$

$$W_Q^h = W_Q^m . W_m^h + W_Q^r . W_r^h \quad (5a) \qquad W_Q^v = W_Q^m . W_m^v + W_Q^r . W_r^v \quad (5b)$$

$$\therefore\ 0.3130 = 0.298W_Q^m + 0.702W_Q^r \qquad \therefore\ 0.3117 = 0.84W_Q^m + 0.16W_Q^r$$

Subtracting $\left[\frac{0.298}{0.84} \cdot (5b)\right]$ from (5a) $\quad 0.2025 = 0.6453W_Q^r$

$$\therefore\ W_Q^r = 0.3138 = 31.38\%$$

By substitution into (5a), $W_Q^m = 0.3111 = 31.11\%$

Restite material should have a concentration close to zero in Q22m - some elements have negative values, i.e. they are more fractionated than apatite. This may mean that some apatite went into the melt, or may simply reflect statistical fluctuations in restite composition between the host rock and the vein (the restite of which forms a fairly small sample).

The calculated melt Q22m consists mainly of quartz, plagioclase and K-feldspar. Its composition is plotted on Figs. 16.12 and 16.13. It lies just inside the plagioclase field, rather than on the cotectic plane, perhaps due to errors in Winkler's field boundaries, or to the specimen not being large enough to be truly representative. Q22m would melt at a slightly lower temperature than Q22a (cf. Fig. 16.14), and could be in equilibrium with unmelted muscovite-plagioclase-quartz residue at pressures above about 6kb - say 5kb to allow for the plagioclase in Q22r being significantly more calcic than An_{30}.

(ii) Trace elements

Calculated trace element contents for restite (Q22r) and melt (Q22m) are presented in Table 16.6. The ratio between the two should represent the bulk solid/liquid distribution coefficient, D, if equilibrium was maintained between melt and restite, and no later re-equilibration took place. To test whether the segregations could be formed by partial melting, this ratio was compared with model D's, calculated from the mesonorm for Q22r and the mineral/melt distribution coefficients (K_D 's) quoted by Hanson (1978) for Ba, Rb, Sr.

Ba Calculated D = 2.2, ratio = 2.05, i.e. very good agreement.

Sr Calculated D = 1.10 or 0.39 (depending on author), ratio = 0.52, bracketed by the calculated values.

Rb Calculated D = 1.11, ratio = 1.82, i.e. the ratio is significantly higher than that expected. Addition of reasonable amounts of additional mica or of some K-feldspar would not raise D to 1.82, so if Q22m is a partial melt, K_D for mica must be higher than the value of 3.26 quoted by Hanson (1978). Fractional or batch melting would not help, since it would deplete Rb still further.

Considering the ratios of these elements, Rb/Sr in the melt decreases because the mica-rich restite has a higher K_D for Rb than for Sr, while K/Ba and K/Rb increase because of the mica's high K_D for Ba and Rb. Sr/Ba in the melt increases, due to mica retaining Ba more strongly than plagioclase does Sr, while Ca/Sr decreases, because Sr is preferentially incorporated in the melt over Ca. Ba/Rb is little changed, since both are strongly retained by mica.

The behaviour of other elements can be considered in the light of their affinities for likely restite minerals (cf. Mason 1966): model K_D's were calculated from Q22r mesonorm and the measured ratio (≡D). Where appropriate, a rough estimate of K_D's in biotite and ores was obtained, by assuming that the ratio K_D^{biot}/K_D^{ore} = (conc. in biot)/(conc. in ore) in the granite quoted by Mason (1966, table 5.4).

Pb enters K minerals; the ratio of 0.62 would imply a K_D of about 2 for micas, which seems reasonable.

Sc goes early into ferromagnesian minerals, especially hornblende - a K_D of 12 for biotite is plausible.

Co goes early into ferromagnesian minerals; K_D's of 6.8 for biotite and 18.7 for magnetite seem reasonable.

Ni has similar properties, and gives a K_D of 5.6 for biotite.

V as V^{3+} replaces Fe^{3+} in magnetite and biotite. Model K_D's of 9.2 for biotite, 185 for magnetite again seem reasonable.

Cr goes into mica and magnetite - K_D's of 1.1 and 18.2 may be a bit low, but all these restitic elements will have a large error in their calculated contents in the melt.

Zn should be retained in biotite and perhaps magnetite. If all the Zn is in magnetite, $K_D = 250$; if in biotite, $K_D = 35$ - values which seem reasonable.

Th is very strongly retained in the restite - it is probably in a minor phase, perhaps zircon, which would give a K_D of about 7000 (sphene, which would be another possibility, is absent from this rock). Σ REE+Th is very large in this rock (>800ppm), so an additional phase such as allanite or monazite may be present, in which case trace element models using Henry's Law cannot be applied.

La, Y and the rare earths are all very strongly depleted in the melt (negative values); they may be contained in some minor phase, and the samples may not be large enough to smooth out statistical fluctuations in its concentration (i.e. the schlieren in Q22a would have a smaller proportion of this phase than the restite in Q22).

In conclusion, the concentrations of most trace elements in Q22r and Q22m are consistent with a partial melting origin for this migmatite.

(iii) Rare earth pattern

Chondrite-normalised rare earth patterns for Q22 and Q22a are plotted in Fig. 16.16. Because of the negative values for Q22m (Table 16.6), D's for the rare earths cannot be calculated. However, D's for all the rare earths would have to be extremely high - they appear to be almost 100% retained in the restite. If partial melting was invoked, the rare earths would have to be contained in restite phases which either had extremely high K_D's, or which were effectively inert and refractory (either due to very low diffusion rates, or to the rare earths being essential components of the mineral, and not released until it melts). Bulk distribution coefficients, calculated from the mesonorm of Q22r, and the K_D's in Hanson (1978), would be far too low (LREE = 0.649, MREE = 0.730, Eu = 0.781, HREE = 1.224). Q22(P) and Q22 (Zr) on Fig. 16.16 were obtained by dividing the values for Q22 by the ratio Q22/Q22a for P and Zr. They indicate the type of curve to be expected in Q22a if the rare earths are entirely contained in restite. On the whole, the MREE and Eu seem to be slightly less depleted than the light and heavy REE. The curve labelled Q22(Σ) was obtained by dividing each element in Q22 by $[\Sigma REE(Q22)/\Sigma REE(Q22a)]$. Most of the disparities with Q22a are probably within errors, although Sm, Eu and Dy may show slight relative enrichment in the melt - possibly indicating the influence of biotite in the restite on the few ppm of REE which are not in a refractory mineral.

Table 16.5 gives some indication of where the rare earths should be if the minerals were in equilibrium at some stage. For all except Eu, 80 - 90% of the total should be in apatite or zircon, which might be refractory, but 50% of the Eu would be in feldspar, and hence available to the melt. In any case, it would be necessary to have some additional restite phase, probably with essential rare earths (e.g. allanite or monazite - not unreasonable, since this rock contains almost 0.1% REE+Th), since the reduction in apatite and zircon from Q22 to Q22a is not sufficient to explain that in the REE.

If the migmatite formed by subsolidus segregation, it might be expected that equilibrium would be achieved between the minerals in the host rock, and those in the vein. From Hanson (1978), bulk D's for Q22 are: LREE = 0.482, MREE = 0.532, Eu = 0.832, HREE = 0.861. The corresponding values for Q22a are 0.160, 0.164, 0.966 and 0.205. *If* these were representative of subsolidus distribution coefficients, the ratios between Q22 and Q22a would be as follows: LREE = 3.01, MREE = 3.24, Eu = 0.86 and HREE = 4.20. The vein should therefore have rather more total rare earths than it does, especially the light REE, and a strong positive Eu anomaly.

In conclusion, the rare earths in this migmatite appear to be contained in an inert restite phase. Regardless of whether partial melting or subsolidus segregation took place, no more than a few percent of the total rare earths can have been involved.

16.3.3 Granitic gneiss and associated pegmatites

(i) Introduction

Three samples were analysed from the Quoich granitic gneiss, which is one of a series of granitic bodies lying near the Glenfinnan/ Loch Eil division boundary (Johnstone 1975). They have been variously interpreted as metasomatised semipelites (Dalziel 1966), intrusive granites (Gould 1965, Pidgeon & Aftalion 1978) or slices of pre-Moine basement (Harris in discussion of Winchester 1974) - cf. sections 2.1, 11.4, 12.3, 17.3. Q11a is a sample of normal granitic gneiss from the spillway. Q13b was taken from the centre of a sinistral shear zone; these zones often develop coarse, pegmatitic textures and veins extend from them into the country rock. Q1 is a part of the coarse, granitoid patch lying in the core of a large open fold at the eastern margin of the granitic gneiss (section 11.3.2). The petrography of these rocks will be discussed in detail in section 17.3. Briefly, they consist of quartz and irregular, corroded oligoclase grains which show reverse zoning at their margins, in a finer-grained, quartz - alkali feldspar matrix. Large grains of alkali feldspar also occur, enclosing the plagioclase and quartz; Q1 is mostly matrix, with many large albite grains, while in Q13b, matrix is relatively minor, and quartz and feldspar are strained, with the latter also fractured. These features would be consistent with the plagioclase and quartz being restite in a quartz - alkali feldspar leucosome, both arising from partial melting of the granitic gneiss.

The granitic gneiss Q11a has a reasonable granitic or ademellitic composition (Appendix 3), with no distinctive major or trace element features. Its normative composition is rather quartz-rich (Table 16.4), so that it plots some distance away from the Q-Or-Ab-An-H_2O cotectic (Figs. 16.12, 16.13), although it lies within the range of compositions determined for the Ardgour granitic gneiss by Gould (1965). The rare earth pattern is also typical of an acid igneous rock (moderately fractionated, with a negative Eu anomaly), and is in fact very similar to that for the Caledonian Foyers granite (Pankhurst 1979). Its origin will be discussed more fully in section 17.3.

Q1 is very albite-rich (Figs. 16.12, 16.13, Table 16.4) with very little K-feldspar; Q13b has somewhat less K-feldspar and quartz than Q11a, and a more anorthitic plagioclase (norm An_{16} vs. An_{11}). However, in detail, the normative compositions of these samples do not fit well with a model of Q1 mainly melt, Q13b mainly restite. A melt derived from Q11a should initially lie close to the minimum melt composition, and should stay on the cotectic until the rock was almost completely melted. The restite would therefore lie further away from the cotectic, in the quartz field, and any restite/melt mixtures should lie between these extremes. Although Q1 lies in the quartz field, and has a low melting temperature (Fig. 16.14), it is much too low in K-feldspar to fit this model, while Q13b lies in the plagioclase field, implying the removal of a very orthoclase-rich melt. The most likely explanation is that the specimens do not represent a series: mainly melt, original granitic gneiss, mainly restite. The samples were fairly small (several hundred grammes for Q1 and Q13b, and about one kilogramme for Q11a), and had to be collected several metres apart, in order that they were well characterised petrologically. A combination of inhomogeneities in the granitic gneiss, and poor sampling of the coarse-grained segregation Q1, might explain the discrepancies. However, their compositions still allow a crude assessment of the likelihood of a partial melting origin.

(ii) Major elements and mesonorms

The felsic components have been discussed above. Relative to Q11a, biotite is reduced in Q1, but similar in Q13b, although it is significantly more magnesian. Muscovite is unchanged in Q13b, but increased in Q1, due in part to sericitic alteration. This may mean that Or in Q1 mesonorm should be larger; a rough estimate of possible error can be obtained by a comparison with the CIPW norm (maximum several percent). From Fig. 16.11, even if the rock had no primary muscovite, the conclusions reached above about melting paths would be unchanged. Magnetite, ilmenite, apatite and zircon are all lower in Q1 and Q13b. In the latter case, this is the reverse of what would be predicted; however, it was apparent in the field that mafic minerals were particularly irregularly distributed in the granitic gneiss.

(iii) Trace elements

Trace elements in Q1 and Q13b were compared with those in Q11a (listed in Appendix 3). If the melting model is correct, Q1 should show an increase in an element where Q13b shows a decrease, and *vice versa*. Equilibrium can be roughly gauged by comparing element ratios with the ratios of appropriate minerals in the mesonorms.

Ba is very severely depleted in Q1. K-feldspar is very low, and biotite is retrogressed to chlorite - possibly Ba has been lost at a late stage. Q13b should have a bulk D of about 1, using the K_D's in Hanson (1978), but fractional or batch melting, which is probably likely in this setting (cf. section 17.3.3) could cause the observed reduction in Ba if D was slightly less than 1. The data are also consistent with equilibrium between the K-bearing phases in Q11a and Q13b, but in Q1, only a four-fold reduction in Ba would have been expected.

Sr increases in Q1 and Q13b - D for Q11a would be about 2.0, for Q13b, 2.5. This is consistent with Q13b being a restite after melting, but not with trace element equilibration between the feldspars of Q11a and Q13b. Q1 should have lower Sr if it is a melt, but it does have a much higher plagioclase content, so subsolidus equilibrium may be indicated.

Rb is reduced in both rocks. D for Q11a or Q13b would be about 0.3, so Q13b could plausibly be a restite; Q1 may be in equilibrium with Q11a, low Rb reflecting the low biotite and K-feldspar content.

For other trace elements, the values of K_D estimated by assuming partial melting in Q22/Q22a (section 16.3.2) were applied.

Pb A K_D of about 2 for K-feldspar and mica would give a D of about 0.5 for Q11a or Q13b. Q13b has the same Pb content as Q11a, perhaps due to a larger K_D for K-feldspar (not present in Q22r), and the presence of trapped melt in Q13b. The lower Pb content of Q1 would suggest restite on a partial melting model, but is consistent with subsolidus equilibrium, given the low K-feldspar and mica content.

Sc, Co, Cr, Ni should have similar contents in Q13b and Q11a and decrease in Q1, but all values are very low (close to detection limit) and so unreliable.

V has similar values in all three rocks, where a slight decrease would be expected in Q1 and Q13b, on either subsolidus or melting models, given their lower magnetite contents.

Th and Zr decrease strongly in Q1 and Q13b, both by similar amounts, possibly due to retention of Th in zircon.

Zn decreases in both Q1 and Q13b, parallelling the reduction in ore content.

On the whole, the trace elements are consistent with Q13b being dominated by restite produced by partial melting of a granitic gneiss similar to Q11a. Q1 appears to be largely in trace element equilibrium with Q11a, which favours a subsolidus origin. However, it may also be due to later re-equilibration - this part of the granitic gneiss suffered particularly severe late retrogression, with plagioclase often replaced by coarse albite-clinozoisite aggregates.

(iv) Rare earth elements

Rare earth patterns for Q1, Q11a and Q13b are shown in Fig. 16.16. Q1 is essentially parallel to Q11a, suggesting that the rare earths were retained in some inert phase. The curves labelled P and Q1(Zr) were obtained by dividing the REE values in Q11a by the ratio Q11a/Q1 for P and Zr (note that Q1 and Q13b have similar P_2O_5 contents). Q1 is largely contained by these curves, indicating that apatite and/or zircon would be suitable inert phases. The HREE are dominated by zircon, and the MREE by apatite, a feature which would be predicted from the K_D values quoted by Hanson (1978).

LREE appear to follow zircon, which would not be expected, but they may be in some additional phase, such as allanite or monazite. Table 16.5 gives the distribution of rare earths between minerals in Q11a if they were simultaneously in equilibrium with a melt (which may be reasonable if the granitic gneiss is in fact a deformed granite). All except Eu are dominated by apatite or zircon. Eu should therefore be available for incorporation into the melt, which would give Q1 a positive Eu anomaly. If partial melting did take place, the rare earths were not distributed as in Table 16.5, although they may have been contained in some additional REE-rich phase.

Straightforward partial melting of Q11a to form Q1 is ruled out by the values of $D_{rock/melt}$, calculated from the mesonorms and the K_D's given in Hanson (1978). These are listed below:

Q1	LREE 0.26	MREE 0.27	Eu 0.91	HREE 0.19
Q11a	LREE 0.36	MREE 0.37	Eu 0.95	HREE 0.29
Q13b	LREE 0.25	MREE 0.26	Eu 1.23	HREE 0.14

A melt having Q11a or Q13b as a restite, in which rare earths were in equilibrium in <u>all</u> phases, should have a similar Eu content, but be strongly enriched in all other REE, especially the HREE, resulting in a much increased negative Eu anomaly. If the ratios of D's are taken as indicative of subsolidus distribution coefficients, then the LREE and Eu contents of Q1 and Q11a should be similar, but the MREE and HREE should be about 50% higher in Q11a.

In conclusion, during the process which formed Q1, the rare earths must have remained in some inert phase; this does not explicitly rule out either partial melting or metamorphic differentiation.

Q13b has a very different REE pattern, dominated by a positive Eu anomaly. Although most values lie between the P and Q13b(Zr) curves (derived in a similar way to those for Q1), the heavy rare earths lie close to apatite, and the light rare earths to zircon, the reverse of what would be expected if the REE pattern was dominated by inert phases. From the ratios of D values given above, subsolidus equilibrium should give Q13b a slight positive Eu anomaly, but the absolute values in Q13b should be much higher, and the HREE should be the most severely depleted, rather than the least.

The positive Eu anomaly and reduction in ΣREE in Q13b certainly suggests that it is a restite after partial melting. In detail, the depletions in REE in the restite are probably greater than would be predicted (note that a melt should have higher REE contents than Q11a). This may be due to fractional or batch melting, which causes a more severe reduction in incompatible elements in the restite, compared with equilibrium melting, although it would not explain the reduction in Eu, which has $D \gg 1$. Some REE may be contained in inert phases, which might be lower in Q13b than in Q11a due to the type of statistical fluctuations discussed above. Alternatively, only the edges of plagioclase grains may be in equilibrium with the melt (note the preservation of reverse zoning in plagioclase).

The broad parallellism between the rare earth patterns of Q11a and Q22 (and LH23a, another Moine pelite - Fig. 16.17) raises the possibility that the granitic gneiss is in part derived by partial melting of Moine-like sediments. D values for Q22r (about 1.2 for HREE, 0.7 for the others - section 16.3.2(iii)) suggest that a melt derived from this rock would have a parallel REE pattern, except for the HREE, which should be somewhat depleted - precisely the relationship with Q11a. A rock with a similar mineralogy to Q22, but with lower ΣREE, and lacking inert REE-rich minerals, would be a suitable starting material.

(v) Conclusions

The REE pattern of Q11a is consistent with derivation by partial melting of a Moine-like metasediment. Q13b appears to be a restite, produced by partial melting of the granitic gneiss in a shear zone, and expulsion of much of the melt. The specimens Q1 and Q13b are probably not large enough to be representative, making it impossible to derive a quantitative model. Q1 is clearly mostly leucosome material - however, its very albitic composition, and trace element equilibrium with Q11a, tend to favour a subsolidus origin. The REE are retained in an inert phase, which could happen in either partial melting or subsolidus segregation, but since the REE in Q13b were not retained in an inert phase, melting is less likely. The very low Ba content, and widespread retrogression, do make a model of partial melting followed by re-equilibration plausible, but on balance, the chemical data favour a sub-solidus origin for Q1.

16.3.4 Late, cross-cutting pegmatite at Kinlochourn

(i) Major elements

LH23 is one of a suite of coarse, cross-cutting pegmatites found in the vicinity of Kinlochourn; they are quite distinct from the regional trondhjemitic lits of the nearby pelitic gneisses. They are syn- to post-$D3_k$ in age, and so probably coeval with the Lochan Coire Shubh pegmatite complex 2 - 3km to the southeast (sections 5.4.2, 10.3). The petrography of LH23 is discussed in section 17.4. LH23a is a sample of the immediately adjacent Morar Division semipelite - a typical example of the type of non-migmatitic pelite discussed in section 16.2 (note the high mica:plagioclase ratio in the mesonorm - Table 16.4).

The intrusive field relations make a local derivation unlikely, although the composition of the two rocks does not rule it out. The anorthite contents of plagioclase are similar (LH23 norm $An_{22.5}$, probe An_{25}; LH23a norm An_{27}, probe $An_{27.5}$); LH23 contains K-feldspar, while LH23a does not. LH23 lies on the Q+Plag+L+V cotectic plane, close to the minimum melt composition (Figs. 16.12, 16.13), and would have a very low melting temperature (around 650°C at 5kb - Fig. 16.14). Above about 3.5kb, it could be in equilibrium with stable quartz+muscovite+plagioclase assemblages. The possible Caledonian P, T estimate of c. 560°C at 4.5kb (section 15.2.3(iv)) would be consistent with intrusion of this pegmatite into somewhat cooler country rocks, hence the sharp contact relationships. The pegmatite was presumably generated at somewhat greater depth and higher pressure.

The pegmatite LH23 is immediately distinguished from the regional migmatitic lits by its high K-feldspar content (Table 16.4). However, unlike other late pegmatites (e.g. Q31a), it has fairly calcic plagioclase. The association of calcic plagioclase and high K-feldspar is also atypical of normal igneous rocks (cf. the granitic gneiss, Q11a, and the felsic porphyrite, Q42). One possible means of producing this feature would be to melt a K-feldspar, quartz, plagioclase-rich semipelite, until most of the plagioclase had been consumed. The melt composition would tend to be stabilised near the cotectic plane, since it would need a large increase in temperature to move it into the quartz field. Suitable starting materials are available in the Moines - e.g. the projections from Q of Q49a and LH11 come close to LH23 (Fig. 16.13). This model is also consistent with the fragments of quartz, mica, garnet-rich restite found in the pegmatites at Lochan Coire Shubh. It will be tested further using the trace element compositions of LH23a, and other metasediments listed in Appendix 3.

(ii) Trace elements

Ba at 21ppm is low, compared with metasediments or other pegmatites - this would be consistent with a micaceous or K-feldspar-rich restite (e.g. D for LH23a is 2, using the K_D values quoted in Hanson (1978)).

Co, Cr, Ni, Sc, Ti, V are all very low compared with the metasediments, and with basic to intermediate igneous rocks (S18b, Q42), but similar to those obtained from other pegmatites.

Pb lies towards the upper end of the range of analyses - D for LH23a would be about 0.7, using a K_D of 2 for mica (cf. section 16.3.2(ii)), resulting in slight enrichment in the melt.

Rb at 104ppm is close to the average for the metasediments, consistent with a calculated D of about 1 for LH23a.

Sr is rather high, except when compared with Q49a. An increase in the Sr content of a melt relative to its source should imply a relatively low feldspar content in the restite (K_D = 2-4, Hanson (1978)) - say a maximum of 10-15% to give D = 0.5.

Th is very low (close to detection limit), but has similar concentrations in the metasediments.

K/Rb, K/Ba are both high (399 and 195 respectively), suggesting the presence of mica in the restite (K-feldspar is ruled out by the high Sr content of the melt).

Ba/Rb is rather low - K_D for Ba in mica is about double that for Rb, so Ba should be preferentially retained in a mica-rich restite.

Rb/Sr is low relative to the metasediments and other pegmatites - mica in the restite reduces Rb/Sr in the melt, while K-feldspar or plagioclase should increase it.

Ca/Y is high, due to retention of Y in micas and/or garnet.

Ca/Sr is low, again suggesting that the residue contains little plagioclase.

Sr/Ba at 1.607 is very high - only the calc-silicate Q48a has a value > 1. This again indicates dominance of mica over feldspars in the restite.

The trace elements are therefore consistent with derivation of LH23 by partial melting of a semipelite, the restite consisting mainly of mica (and quartz), with no more than about 10% feldspars.

(iii) Rare earth elements

The rare earth pattern for LH23a (Fig. 16.17) is similar to that for the pelite Q22 (Fig. 16.16), but with much lower ΣREE. The pegmatite LH23 has a very different pattern, with a strong positive Eu anomaly, and very low ΣREE, especially the HREE. Qualitatively, the positive Eu anomaly in the melt indicates little or no feldspar in a partial melting restite, especially if the source rock had a negative Eu anomaly like Q22; however, it might suggest subsolidus equilibrium with the adjacent metasediments. As before, this can be tested by assuming that the ratios of D values are representative of high temperature subsolidus distribution coefficients (cf. section 16.3.2(iii)). Bulk distribution coefficients are (Hanson 1978):

LH23 LREE 0.132 MREE 0.124 Eu 0.734 HREE 0.095
LH23a LREE 0.256 MREE 0.261 Eu 0.501 HREE 0.302

The concentration of LREE and MREE should be about twice as great in LH23a, and those of Eu and HREE two-thirds, and three times as great respectively. In fact, the HREE are about 2.5 times as great, Eu is unchanged, while the LREE and MREE show the largest difference (four and three times respectively). The data are therefore not particularly consistent with subsolidus equilibrium.

LH23 is unlikely to be a partial melt of LH23a, since given the D values above, such a melt should have much higher total rare earths, with a very pronounced negative Eu anomaly. However, it may be that some of the rare earths were retained in inert phases. If they were distributed as in Table 16.5, then for all but HREE, a large proportion would be in feldspar and mica, and presumably available for incorporation into the melt. It would be necessary for all but Eu to be contained in some other inert phase - this seems unlikely, since in such a situation (Q22/Q22a) all the REE are retained. Assuming that a metasedimentary source would have a fairly flat REE curve like LH23a, the restite would have to contain very little feldspar, but include minerals such as apatite, garnet or hornblende to give a positive Eu anomaly in the melt. The very low values for the HREE in LH23 suggest the influence of garnet or zircon, which have higher K_D's for these elements.

The rare earths are therefore consistent with the idea that this late pegmatite was produced by a large degree of partial melting at depth of a K-feldspar - bearing semipelite or micaceous psammite.

16.3.5 Other rocks

(i) Q31a is a very late, coarse-grained (several centimetres) K-feldspar/albite/muscovite - bearing pegmatite. It is one of a suite of dykes, which have sharp, 015°-trending contacts and cut the quartzites in the spillway at Loch Quoich (section 11.1). It is very quartz-rich (Table 16.4; Figs. 16.12, 16.13), and lies some distance from the minimum melt composition, but the 1-2 kg sample may not be representative of such a coarse-grained rock. Trace element contents - fairly high Ba (almost 1000ppm), low values for REE and elements associated with mafic minerals, moderate Rb, Sr (130 and 185 respectively), K/Rb = 300 - are all typical of a very acid igneous rock.

(ii) Q42 is a felsic porphyrite from Loch Quoich. This is a medium-grained intrusive rock containing abundant zoned feldspar phenocrysts; it is slightly metamorphosed, the quartzofeldspathic matrix being recrystallised (cf. section 11.1; Smith 1979). It has a broadly granodioritic or quartz-dioritic major and trace element composition (Appendix 3 - but note the high Sr content, reflecting the presence of plagioclase phenocrysts). The REE pattern (Fig. 16.17) is very similar to those from the Strontian and Foyers granodiorites (Pankhurst 1979), but slightly less fractionated due to lower values for the LREE (70 - 80 vs. 100). The lack of a Eu anomaly is explained by compensation of a probable negative anomaly in the matrix by a positive anomaly in the plagioclase phenocrysts. The normative composition (50% plagioclase - Table 16.4) lies well away from the cotectic, in the plagioclase field (Figs. 16.12, 16.13).

(iii) Q49a is a very feldspathic, micaceous migmatitic psammite from the quarry by Loch Quoich (section 11.1). It has a high K-feldspar content, but fairly calcic plagioclase (An_{29} - cf. Table 16.4); a relatively low Rb/Sr ratio and high K/Rb reflects its high K-feldspar content. It lies close to the Q+Plag+L+V cotectic surface (Figs. 16.12, 16.13), and could melt at a relatively low temperature (<660°C at 7kb - Fig. 16.14). It may be a partial melt migmatite, and would also make a suitable source rock for late pegmatites such as LH23.

(iv) S12 is a sample of the highly-foliated material at the margin of the Strontian granite (section 13.2). It appears to intrude and brecciate quartzites and amphibolites, but in the field, this was seen to be due to extreme ductility contrasts, rather than to its having been molten. It was analysed to confirm that it is not part of the deformed margin of the granite. Its composition is clearly semipelitic rather than igneous - note the combination of high SiO_2 with relatively high CaO/Na_2O and ΣFe as Fe_2O_3. Trace element contents are similar to other semipelites and micaceous psammites. It has over 40% normative quartz (Table 16.4), 10%

K-feldspar associated with An_{31} plagioclase and 17% mica. It plots well away from the cotectic surface, in the quartz field (Figs. 16.12, 16.13), but could undergo partial melting at a relatively low temperature if $P_{H_2O} = P_{load}$ (at the Kfs+Plag+Q+Ms+V = L curve on Fig. 16.14). However, P_{H_2O} in the inner aureole of an I-type granite intruded into previously metamorphosed rocks could well be less than P_{load} (e.g. Ashworth & Chinner (1978) estimate $P_{H_2O} = 0.6P_{load}$). In that case, melting temperatures would be increased (Fig 16.14). Tyler & Ashworth (1982) have suggested that this rock was partially melted during forceful intrusion of the granite, with an intergranular melt film reducing its viscosity, but that discrete segregations did not form, and textural evidence for melting was eliminated by recrystallisation.

(v) LH11 was collected from the coarse augen gneiss in the Reidh Psammite, adjacent to the hornblendic Lewisian inlier at Kinlochourn (section 5.1.3). This lithology occurs widely at the base of the Glenfinnan Division, and is spatially associated with the Lewisian inliers. It could be a basal Glenfinnan Division rock, or a slice of more acid Lewisian basement. Its composition is clearly metasedimentary - note particularly the high SiO_2/Al_2O_3, CaO/Na_2O, for a rock whose K_2O content would suggest a granite. These features are also reflected in the norm (Table 16.4), which plots well away from any igneous composition, in the quartz field (Figs. 16.12, 16.13). If it is a basal Glenfinnan Division lithology, it might be expected to have relatively high Ti, Zr and Cr, Ni etc. This is not clear from the analyses (Appendix 3), although there is a shortage of other psammites for comparison. In any case, if it was derived from a dominantly granitic terrain of Laxfordian aspect (suggested by the high K-feldspar content), it might not inherit such distinctive features.

CHAPTER 17

Origin of migmatites

17.1 Regional "trondhjemitic" migmatites

These rocks form the bulk of the migmatites within the Moines, and are typically coarse pelitic gneisses of lit-par-lit type (stromatic type of Mehnert 1968). They are most common in the Glenfinnan Division, which is of generally higher grade than the Morar and Loch Eil divisions (section 15.3.2), and which contains a greater proportion of readily-migmatised lithologies (section 16.2.5). The descriptions given below are based on several hundred samples collected from the areas discussed in Chapters 3 to 13.

Where undeformed, the migmatites contain coarse-grained, irregular, quartzofeldspathic pods and lenses, lying in an earlier strong tectonic fabric. In most areas they have suffered intense subsequent deformation; this generally begins with the Sgurr Beag Slide and correlative structures, with further deformation resulting from later tectonic events (Plates 5.1(a), 8.1(a), 17.1(a)-(c)). In slide zones, feldspar is augened and wrapped by quartz ribbons; elsewhere, perhaps due to lower strain rate and different P,T conditions, the contrast between these minerals is less marked, and the whole lit is recrystallised to a relatively fine grain size.

From section 14.4, it is likely that this migmatisation was Precambrian, and the reworking Caledonian. Cross-cutting pegmatites related to the regional migmatites are extremely rare, and no evidence for intrusion of the leucosome (e.g. rotated blocks of palaesome) was observed. Those large pegmatites which do occur (e.g. a post-D2, pre-D3 (slide) pegmatite in Knoydart, Chapter 4) could well be members of the Morarian suite, and may represent a distinct, post-migmatisation, pre-Caledonian event - D. Powell (verb. comm.) has found such veins cutting the garnet-grade Morar Division rocks of Moidart (cf. section 14.4).

In thin section, the leucosome of the trondhjemitic regional migmatites consists of quartz and plagioclase (usually An_{25-30}), with minor amounts of apatite, zircon and (sometimes) garnet. Coarse-grained biotite-muscovite-garnet screens (former selvedges?) occur within the larger veins. In undeformed migmatites, plagioclase is tabular and subidioblastic with 0.5-2cm grain size (Plate 5.2(g)), and quartz is very coarse (several centimetres - but since it recrystallises very easily, large original grains are recognised by the presence of areas with very similar crystallographic orientation). With deformation, the quartz is quickly recrystallised to a few millimetres grain size and plagioclase is also affected (Plate 5.2(f)). In extreme cases, all grains are reduced to 1-2mm, with quartz defining a shape fabric (Plate 13.1(d)). K-feldspar is generally absent from the leucosome, except in those rare cases where it is also present in the host rock (usually in the more semipelitic compositions). The plagioclase is occasionally albitic, due to later retrogression to albite+clinozoisite (e.g. on Mull, section 15.2.3(v)) or to a very sodic bulk composition (e.g. on Ben Klibreck, section 9.4.3).

The melanosome consists of quartz, muscovite, biotite, garnet, plagioclase and ilmenite; accessory apatite, zircon and monazite were identified, and sphene and/or haematite were often produced by retrogression (biotite remaining stable). The texture is typically schistose (Plates 8.3(b), 17.1(b)), with the variation in grain size from 0.1-1mm (quartz) and 0.2-2mm (biotite) being due in part to later deformation. In less deformed rocks, the quartzofeldspathic portion of the melanosome forms small lenticles, like miniature versions of the migmatitic lits (Plate 8.3(a)). A coarser-grained (c. 5mm) biotite-rich selvedge is often, but not invariably, present at the margin of the leucosome (Plate 17.1 (b), (c)).

Aluminosilicates are rare, except at Mull and Loch Ruthven, where they are severely affected by later Caledonian thermal and regional retrogressive metamorphism, and K-feldspar/sillimanite or K-feldspar /kyanite associations were not recorded in the regional migmatites, except in the aureoles of Caledonian intrusions (e.g. the Strontian granite, Chapter 13). Fibrolite is widespread though not common, but is texturally very late, and may be Caledonian. A regional K-feldspar-sillimanite zone occurs in Ardgour (Stoker 1980), although migmatites in this area (north of Glen Tarbert, [NM 922622]) are texturally quite distinct from the "normal" regional migmatites (abundant evidence of extreme viscosity contrasts, if not partial melting; numerous cross-cutting pegmatites and aplites; semipelites apparently intruding quartzites and amphibolites - Plates 17.1(d), 17.2(a),(b)).

The trondhjemitic migmatites are only found in the kyanite and sillimanite (-muscovite) zones. Due to the generally psammitic nature of the Morar Division, and the cutting out of part of the metamorphic sequence by the Sgurr Beag Slide, lithologies suitable for migmatisation do not reappear down-grade until the garnet zone. There is therefore some doubt as to the lowest grade at which these migmatites could form.

The four general mechanisms for the production of migmatites are shown in Table 17.1, together with criteria for distinguishing between them (cf. Yardley 1978).

An open system origin for the migmatites is unlikely. Virtually none of the petrographic criteria for igneous injection are met, and in any case intrusion of numerous tiny, often blind veins into a huge volume of rock, with no sign of a suitable pluton in the area, is highly improbable. Steveson (1968), in a rigorous statistical study, showed that metasomatism in the migmatised Morar Division must be very minor or absent entirely. The mafic selvedges have a complementary composition to the leucosome, rather than one intermediate between vein and palaeosome, and the detailed nature of the migmatite/non-migmatite interbanding favours control of migmatisation by original bulk composition (section 16.2.1). From section 16.2.3, migmatites <u>within the Glenfinnan Division</u> have some chemical features which might be taken to indicate metasomatism (high Na_2O, restricted range of bulk compositions, including oxidation ratio - but note the lack of any distinctive trace element features), but this is relative to the interbanded non-migmatites, <u>not</u> to low-grade equivalents. The Garnetiferous Pelite of the low-grade Morar Division is lithologically equivalent to the typical Glenfinnan Division migmatite, and has similar chemical features (sections 16.1, 16.2). The distinctive chemistry of the migmatites is therefore interpreted as a primary sedimentary feature, and as a factor which controls the process of migmatisation, rather than a consequence of it.

The choice therefore lies between partial melting and metamorphic segregation. From section 16.2, the trondhjemitic migmatites lack features such as a higher Na/Ca ratio than associated non-migmatites, or a high K-feldspar content, which would have favoured partial melting. Geothermometry indicates temperatures below those at which a K-feldspar - free pelite would begin to melt (section 15.3.1), and the leucosome composition is incompatible with partial melting of the melanosome (no K-feldspar, no differentiation in plagioclase composition). The conclusion must be that the trondhjemitic regional migmatites formed by subsolidus metamorphic segregation.

Models for subsolidus migmatisation are less well constrained than those for partial melt migmatites. Clearly the replacement of many small grains by fewer large, equant grains is favoured by the minimisation of interfacial energy, as is the segregation of quartz and feldspar from mica and garnet (grain boundaries between these groups - e.g. plagioclase/biotite - have generally higher energy than those within each group - e.g. quartz/plagioclase). The basic mechanism must be solution of the quartzofeldspathic components, transport by diffusion through a water-rich fluid, and deposition in the leucosome. This, however, does not approach the question of why some rocks are more susceptible to this process than others. Two end-member explanations exist for the non-migmatisation of certain rocks. Either their composition was such that migmatisation could not take place (i.e. the mechanism of textural development was different in migmatites and non-migmatites), or the process was so slow in the non-migmatites that it would have taken a geologically unreasonable length of time to reach completion.

Olsen (1977) has proposed a model of the first type, in which local partial melting sets up activity gradients which drive the diffusion of additional quartzofeldspathic material to the developing leucosome, whose final composition may be (say) 20% melt, 80% metamorphic segregation. Only rocks with a low-melting composition would be affected, even though the final leucosome may lie far from a cotectic composition. However, the regional migmatites show no correlation with a low-melting composition, and even on Olsen's model, the leucosome should have several percent K-feldspar.

Steveson (1968) developed a model whereby migmatisation was controlled by Na^+/H^+ and K^+/H^+ ratios in the fluid phase, with deposition of feldspars in low pressure sites. On the scale of a bed, these ratios should be controlled by the rock type, but a tectonic factor could also affect migmatisation; a feature favoured by the field evidence. Since the migmatites are MP2, the segregations perhaps formed in the S2 fabric following release of stress at the end of D2. There would be some limiting value of fluid composition, below which no deposition could take place, and it might be expected that deposition of plagioclase would be more extensive in a rock with a high Na^+/H^+ ratio. However, this model does not fit particularly well with the bulk chemical controls on migmatisation. Although high Na_2O favours migmatisation, this is linked to the plagioclase content of the rock, rather than the albite content of the plagioclase (since the Na_2O/CaO ratio is not significantly greater in the migmatites - section 16.2). At equilibrium, the activity of Na^+ or of albite component in the fluid is controlled by the <u>mineral</u> composition, not the rock composition (assuming that the system was closed on the scale of a bed, as suggested by Steveson's (1968) results). Therefore perhaps equilibrium was not maintained.

Some supporting evidence for a rate-controlled model can be obtained from lower grade parts of the Moines. Among the garnet -grade rocks of Mallaig, the Garnetiferous Pelite, which is compositionally equivalent to the readily-migmatised Glenfinnan Division pelites, is significantly coarser-grained than the other Morar Division pelites and semipelites; the coarsening is clearly post-tectonic, and not just inherited from the sedimentary grain-size. It contains abundant small quartzofeldspathic veins and lenticles, and is thus similar in texture to the high-grade non-migmatites. A mechanism-controlled model would require that different mechanisms produced the same texture in low-grade garnetiferous pelite and high-grade non-migmatite, or that there was a change of mechanism from low-grade garnetiferous pelite to high-grade migmatite. The latter suggestion is at variance with the close similarities between these rocks; the garnetiferous pelite is similar to the lenticular-segregated matrix of partially-migmatised high-grade rocks, which is identical in all but grain-size to the coarser banding of highly-migmatised rocks.

As an example of how a rate-controlled mechanism might operate, suppose that migmatisation progressed relatively slowly in geological terms, and the rate-controlling factor has an Arrhenius-type temperature dependence (rate constant = $Ae^{-E/RT}$, where A has the same units as the rate constant and is constant for a particular reaction, E is the activation energy, R the gas constant and T the absolute temperature); also that the duration of the garnet-grade event at Mallaig was the same as that of the sillimanite-grade event in the migmatites, with the temperature being 100°C lower. For most chemical reactions, a 10°C rise in temperature results in a doubling of the reaction rate. A rock can be described as a migmatite when a significant number of quartzofeldspathic veins have been produced; in effect, when the reaction has reached a certain percentage of completion. Call a typical migmatite/garnetiferous pelite lithology Rock A, and a typical non-migmatite Rock B. The degrees of segregation of Rocks A and B are similar where there was a 100°C difference in metamorphic temperatures. Therefore, at the same temperature, the process would be 1024 (2^{10}) times as fast in Rock A as in Rock B. If it took several million years to migmatise Rock A, it would take several billion years to migmatise Rock B - i.e. in a normal orogenic event, it would never be migmatised. Alternative, Rock B could be migmatised in a reasonable time if temperatures were 100°C higher than in the sillimanite-grade Moines. However, at such a temperature (> 700°C), there would certainly be a change of mechanism (partial melting would become important) and so a simple extrapolation up-temperature could not be applied. Thus in a given area, a range of lithologies with a range of rates of migmatisation would result in a mixture of migmatites and non-migmatites.

There are two main ways in which the plagioclase content could control the rate of growth of segregations. If the solution process was slow, a very large activity gradient would exist over the first few microns around each grain, with a very gentle gradient and slow diffusion throughout the rock, and the growth of segregations would be controlled by the solution rate rather than the activity gradient. Clearly a large plagioclase content would present a large surface area, enabling more rapid solution. Alternatively, the diffusion rate may itself be slow, regardless of the activity gradient - then a large plagioclase content would result in statistically shorter average distances across which material must diffuse to produce a given size of segragation.

Migmatisation in these rocks is therefore the culmination of processes which were already in operation at lower grade, causing anomalously rapid grain growth in appropriate lithologies. The most highly-migmatised rocks (consisting entirely of neosome - vein and selvedge - e.g. Plate 6.5(a), part of Plate 17.1(c)) can be regarded has having reached textural equilibrium, i.e. any further diffusion would not significantly alter their appearance; the typical migmatites (with about 50% coarse-grained but non-segregated palaeosome - e.g. Plate 6.4(c)) are part-way there, while the non-migmatities (e.g. Plate 6.4(f)) had barely started the process when the rocks began to cool down.

17.2 K-feldspar-bearing migmatitic pelite

On either side of the Quoich granitic gneiss (Chapter 11), distinctive K-feldspar - bearing pelitic bands occur within the Loch Eil Division. These are coarse pelitic gneisses, which carry concordant pegmatitic veins, several centimetres thick and several metres long, in a matrix which itself contains numerous quartzo-feldspathic lenticles. The western outcrop is narrow, and succeeded by quartzites and then feldspathic psammites. To the east of the granitic gneiss, a thicker pelitic unit occurs; this includes non-migmatitic quartzites and semipelites, and normal "trondjhemitic" migmatitic pelitic gneisses, as well as the K-feldspar - bearing variety. The K-feldspar - bearing pelites can be distinguished in the field by their very low garnet and high biotite contents, by their highly migmatitic nature, and by the fact that their pegmatitic <u>lits</u> seem on the whole less deformed than those in the "trondhjemitic" migmatites.

In thin section, the <u>lits</u> consist of quartz, plagioclase (An_{20-25}) and K-feldspar. <u>Schlieren</u>, containing biotite, zircon, apatite and ilmenite, with rare muscovite and garnet, probably represent fragments of the mafic selvedge, but some garnet appears to have crystallised in the lit, and some muscovite is certainly secondary. Plagioclase and K-feldspar are coarse-grained (> 1cm), the former being subidioblastic, while the latter has irregular, lobate grain boundaries, and may have crystallised somewhat later. The feldspars are very strongly perthitic or anthiperthitic (with 30-40% of the minor phase). K-feldspar has patch perthites with plagioclase inclusions up to 1mm in diameter, while in plagioclase, the inclusions are smaller and aligned in the plagioclase cleavages. Typically, the lit contains more plagioclase than K-feldspar, and has both homogeneous plagioclase and antiperthites, with no separate K-feldspar grains (e.g. Q22 - 56174). In more K-feldspar - rich rocks, homogeneous plagioclase and perthite was recorded, or antiperthite only, and in extreme cases, the lit consists entirely of quartz and perthite. In all rocks, each individual phase (host and inclusion) is in optical continuity. Quartz is generally much finer-grained (1-2mm) and polygonal, and although internally unstrained, has probably recrystallised during deformation. These veins are often boundinaged, and have certainly been affected by the "conjugate" (late $D2_q$) and $D3_q$ structures (cf. section 11.2.1), but were never seen to be deformed by the $F2_q$ isoclinal folds and shear zones. In the more deformed rocks, e.g. Q23 (54665) from the western contact of the granitic gneiss, some antiperthites are recrystallised to plagioclase/K-feldspar aggregates with 1mm grain-size and aspect ratios of about 3:1. In the same rock, relict coarse plagioclases retain their antiperthitic texture.

The lit is invariably separated from the palaeosome by a mafic selvedge, which consists largely of biotite, with apatite, zircon, opaques, minor garnet and muscovite. This is usually several millimetres thick, and has micas up to 1cm in length. The host rock is very coarsely schistose (or finely gneissose), quartzofeldspathic lenticles a few millimetres by a few centimetres being wrapped by a mica-rich foliated matrix. The lenticles contain quartz and perthitic or antiperthitic feldspars, like those of the lits, but are much finer-grained (1-3mm - cf. Plate 17.2(c)). A much greater proportion of normal twinned plagioclase is present, of An_{30-35} composition; this appears to be associated with the matrix rather than the lenticles, although the boundaries between these are diffuse in thin section. The fine grain-size and lack of continuity of the lenticles is probably original, rather than the result of deformation and transposition as in some of the trondhjemitic migmatities, since the antiperthites have not been recrystallised.

Clearly, the migmatisation process in these rocks produced both coarse pegmatitic veins and smaller quartzofeldspathic lenticles; the larger-scale redistribution of material required for the former gave rise to mafic selvedges, suggesting the operation of an isochemical mechanism (cf. Table 17.1). The chemistry of sample Q22, from the western outcrop, was investigated in section 16.3.2, and was found to be consistent with a partial melting origin for this migmatite. The differentiation in plagioclase composition between host and lit and the low-melting composition of the latter are particularly compelling, although no intrusive features were observed, perhaps due to the very strong earlier foliation carried by these rocks.

The mass balance model of section 16.3.2 suggested that the vein contained some restite, and the host some melt. The quartzofeldspathic lenticles in the host could represent melt which had not migrated to form a large vein. In the vein, some of the non-perthitic plagioclase could represent xenocrysts of restite plagioclase. Alternatively, all the restite could be contained in the mafic schlieren, and the plagioclase could have crystallised early out of the melt (note that the calculated melt for Q22 lies in the plagioclase field). The antiperthites would have been produced later, by simultaneous crystallisation of plagioclase and K-feldspar. Those antiperthite-bearing rocks which lack separate plagioclase crystals in their lits were probably more K-feldspar - rich, and produced melts which lay close to the cotectic, while the perthite-bearing rocks must have had a still higher K-feldspar/plagioclase ratio. Simultaneous crystallisation is probably a better explanation for the perthitic textures than exsolution from hypersolvus feldspars. The temperatures required for the latter would be very high, bearing in mind that the Na-phase contains considerable anorthite component, and that some compositions would lie virtually on the solvus crest.

In addition, the coarseness, and poor crystallographic form but good crystallographic orientation of each phase do not favour exsolution. Furthermore, some feldspars show swap textures (K-feldspar and plagioclase grains, linked by a zone several millimetres wide which grades from perthite at one side to antiperthite at the other), which are more readily explained by simultaneous crystallisation. This process could take place during either subsolidus segregation or crystallisation from a melt, but is probably more likely during the latter, when the crystals can grow into a free (liquid) environment, perhaps nucleating on one another, or sticking together while still small grains.

The question then arises: what age are these migmatites, and what is their relationship to the regional trondhjemitic lits? The K-feldspar - bearing lits must be relatively late in the structural sequence - they show slight or moderate deformation, and certainly nothing consistent with the intense $S2_q$ fabric (containing rare isoclinal folds) which characterises all the rocks in the spillway section, including the trondhjemitic migmatites. The small lenticles also appear to have segregated in $S2_q$, rather than having been augened by it. P,T estimates for the pre-$D2_q$ metamorphism in this part of the Glenfinnan Division (synchronous with trondhjemitic migmatisation) are below even the granite solidus (590°C at 5.5-6kb). However, there are strong indications that post-$D2_q$ partial melting took place in the adjacent granitic gneiss, particularly in the shear zones. It is likely that migmatisation in these rocks post-dated the main $D2_q$ tectonic event, but pre-dated shear zone development, and hence was Caledonian (cf. section 14.4). The relatively low strain in the lits suggests that they did not fully crystallise until after the shortening (normal to bedding) which produced the shear zones. The rocks may already have been segregated during the Precambrian event, but if so, the textures have been completely obliterated by this Caledonian migmatisation.

One intriguing possibility, suggested by the presence of these rocks on either side of the granitic gneiss, was that they represented a cryptic aureole to the original granite (either as hypersolvus feldspars in the aureole of a shallow granite, or as veins of granite penetrating the metasediments). The time relations certainly oppose this suggestion (the granite was emplaced prior to $D2_q$, and probably prior to $D1_q$), and arguments against a hypersolvus origin for the perthites have already been advanced. The lits have distinctly more calcic plagioclase than the typical granitic gneiss, and a compositional variation more easily explained in terms of locally-derived migmatites than in late pegmatitic veins derived from a granite. The selvedge is complementary to the leucosome (cf. Table 17.1), rather than of intermediate composition. Although the rare earth patterns for Q22a (lit) and Q11a (granitic gneiss) are broadly similar, there are divergences at the heavy end, and other trace elements are also different (cf. sections 16.3.2 and 16.3.3).

These distinctive migmatites are therefore best explained as being due to the combination of unusually high Caledonian temperatures in the Quoich dam area (also reflected in the remobilisation of the granitic gneiss - section 17.3.2), and a Loch Eil division pelite with a bulk composition quite different from those of the typical Glenfinnan Division pelites. The original (sill-like?) granite may have chosen this pelite, in a thick psammitic sequence, as a weak horizon in which to intrude.

17.3 West Highland granitic gneiss

17.3.1 Origin of the granitic gneiss

(i) Introduction

The West Highland granitic gneiss extends in a discontinuous belt from Ardgour to Fort Augustus (Fig. 17.1; Johnstone 1975). The large body at Ardgour lies close to the Glenfinnan/Loch Eil division boundary, and a series of smaller concordant bodies run sub-parallel to this boundary as far as Loch Quoich. There, the granitic gneiss runs eastwards until it reaches the Great Glen fault. At Ardgour, it is at its deepest structural and stratigraphic location, and also in its highest grade environment (sillimanite zone); as it climbs into the Loch Eil Division, the grade of metamorphism in the country rocks declines (kyanite? at Loch Quoich, garnet? at Fort Augustus).

Various theories have been propounded for its origin. Harry (1953) suggested that the trondhjemitic regional migmatites (section

17.1) were produced by Na metasomatism, and the Ardgour gneiss by subsequent K metasomatism (based on their compositional differences from a handful of non-migmatitic pelites). Dalziel (1966) accepted that the regional migmatites formed isochemically, but maintained that the Ardgour gneiss was produced by K metasomatism in a fundamental shear zone corresponding to the Loch Quoich Line, which separates a mobile infrastructure of Glenfinnan Division pelites from overlying Loch Eil Division psammites. The presence of rounded (sedimentary) zircons and an apparent lateral passage into metasediments round a fold core were considered particularly important.

Gould (1965) noted the typically igneous chemical composition of the Ardgour gneiss, but came to no firm conclusions regarding its origin. He found a broad scatter around granite-adamellite compositions, with the average lying on the Q side of the Q-Plag-Or cotectic plane, near to the minimum melt composition at several kilobars pressure. There was a north-south variation in Na, Ca, K, Si, Fe, Mg, Mn, Ti, Zr, P content, the southern outcrops being slightly more differentiated. He suggested that this was part of a calc-alkaline trend, and finding most of his analyses to be Al-undersaturated, postulated a mixture of Moine- and Lewisian-derived melts. He accepted Dalziel's arguments for the production of granitic gneiss from metasediment, and suggested that this was due to injection of a silicate liquid rather than metasomatism by an aqueous fluid.

Harris (in discussion of Winchester 1974) suggested that the Quoich gneiss could be a slice of pre-Moine basement (Lewisian or Grenville), tectonically emplaced into the sequence. Pidgeon & Aftalion (1978) found evidence in the Ardgour zircons for a pre-Caledonian history, and pointed out that both "igneous" and "sedimentary" zircons were present, and that the latter type was not uncommon in other, undoubtedly igneous Older Granites. Aftalion and van Breemen (1980) performed further zircon studies, the results of which will be discussed later.

Most of the arguments for metasomatism can no longer be considered valid. Harvey's (1953) were statistically meaningless; a series non-migmatitic pelite, migmatitic pelite, granitic gneiss would inevitably produce the chemical differences discussed, and additional evidence that this series describes a geological process would have to be adduced. Dalziel's (1966) zircon evidence has now been discredited, and his map evidence could equally be explained by a granite with somewhat discordant contacts: some tectonic sliding would be required to explain the full outcrop pattern, but this is *necessary even on* Dalziel's model. If metasomatism produced a large body like the Ardgour gneiss, it would surely result in a fairly wide, diffuse zone of partially altered material. Although the granitic gneiss is interbanded with normal metasediments near its margins, rocks of intermediate composition are restricted to a few small psammitic fragments which could easily be xenoliths, partially metasomatised by the enclosing intrusive granite. Gould admitted that the granitic gneiss has a typical igneous composition, and that if it was a magma/sediment mixture, the magmatic portion would have an unusually Al-undersaturated composition.

The time relations of migmatisation presented by Harry (1953) and Dalziel (1966) are incorrect, due to their failure to recognise transposition of early migmatitic lits into later strong fabrics, and to distinguish between the products of regional migmatisation and the much later Caledonian intrusive pegmatites, which may be either albite-rich or K-feldspar - rich. In fact, as will be shown below, the granitic gneiss was deformed during D1, and both it and the metasediments underwent migmatisation or segregation post-D1, pre-D2. The Loch Quoich Line therefore has nothing to do with the formation of the granitic gneiss. Rather than being a D2 shear zone, it is a zone across which D3 (late Caledonian) folds tighten (cf. sections 11.2, 11.4, 12.3); the granitic gneiss outcrop is folded by it to the north of Loch Quoich, and both it and the regional migmatites were in existence long before this structure formed.

At Ardgour, the gneiss shows considerable interbanding with the adjacent metasediments, with no signs of exceptionally high strain. At Loch Quoich, although the gneiss and its surrounding metasediments lie in an area of generally high strain, they certainly lack the intense fabrics associated with major structures such as the Sgurr Beag Slide, or the margins of Lewisian inliers in Sutherland (Chapter 9). Thus although an igneous origin has not been explicitly proven, the alternatives can no longer be sustained, and it is the easiest way to produce a large, homogeneous body of "granitic" composition.

(ii) Petrography

The typical granitic gneiss is a medium-grained (1-2mm), equi-granular rock, with subequal amounts of quartz, oligoclase and K-feldspar. Biotite, up to 10%, in thin, discontinuous laminae, defines a strong foliation in which lie coarser pegmatites. Minor muscovite, garnet, ilmenite, zircon, apatite, monazite and pyrite are present. The Quoich gneiss is somewhat more acid than that at Ardgour (plagioclase An_{10-15} vs. An_{20-25}, significantly less biotite present), possibly indicating an extension of Gould's (1965) chemical variation.

The dominant tectonic fabric in the granitic gneiss is equivalent to local S2 in the metasediments, the first fabric to deform the lits in the regional migmatites. This strong gneissic fabric also deforms and transposes the early pegmatites in the granitic gneiss (Plates 11.2(a),(b); 12.1(a),(b); 17.2(d)). A distinction can be made between banding caused by thick pegmatites, often carrying a strong internal fabric parallel to S2 (Plate 11.1(d)), and by transposition of thinner mica-free lits (Plates 11.2(a),(b); 17.2(d)), and the fabric in homogeneous granitic gneiss, defined only by thin, impersistent biotite laminae. At Loch Quoich, and over much of the Glenfinnan section, the fabric observed is a composite of all three.

At Loch Quoich, traces of an earlier gneissic fabric were seen in rare $F2_q$ fold cores, where it may be slightly oblique to the coarse pegmatitic lits. This earlier fabric is well preserved in the core of a major $F2_g$ fold at Glenfinnan (section 12.2), where it is a deformation fabric, with no associated segregation or veining (Plate 17.3(a)). At least the thicker, pegmatitic veins cut this S1 fabric; their high angle to S1 and lack of internal S1 fabrics (but presence of weak S2 fabrics) suggest an MP1 age, coeval with regional migmatisation. In this low strain zone, they are a few centimetres thick, less than one metre long, and often lenticular, with a prominent biotite selvedge (Plate 17.3(a),(b)). Any possibility that these veins were produced by injection or metasomatism, with the selvedges being relict metasediment is clearly ruled out by these relationships. The non-segregated part of Plate 17.3(a) is representative of the pre-migmatisation form of the granitic gneiss, and could reasonably be interpreted as a deformed granite. Judging by the augened feldspars, it was not particularly porphyritic, and had a grain size of a few millimetres to one centimetre.

The early pegmatitic lits at Glenfinnan contain very coarse (several centimetres) quartz, albitic plagioclase and K-feldspar, e.g. G2a (58145). quartz is recrystallised to grains measuring 1mm by 3mm and shows strong undulose extinction, which defines an $S2_g$ shape fabric. The feldspar is extended and recrystallised to aggregates of 0.2mm polygonal grains. Some of the thicker veins are zoned, from pyrite at the centre, out of which lath-like crystals of An_{12} plagioclase showing good Carlsbad-Albite twinning extend into the quartz-rich margin of the vein, where recrystallisation is more marked than near the centre. The host granitic gneiss may have slightly more calcic plagioclase (An_{15-20} ?), but either a partial melting or subsolidus origin is possible for these veins (the K-feldspar - rich granitic gneiss could melt at temperatures well below the solidus for the surrounding K-feldspar - free pelites). The Quoich veins are finer-grained (several millimetres) and have equant textures, perhaps due to the greater Caledonian deformation and recrystallisation in this area.

(iii) Conclusions

The precursor of the granitic gneiss was intruded into the Glenfinnan and Loch Eil divisions as an igneous body prior to the climax of the D1 tectonic event. Since D1 in this area probably covers the entire pre-Caledonian history of these rocks, it probably consisted of several tectonic phases (equivalent to D1 and D2 of Mallaig, but indistinguishable due to high Caledonian strain), so the granite could have been syntectonic rather than pretectonic, a Precambrian analogue of the Caledonian Older Granites. The lack of D1 cataclasis, or segregation, suggests neither very low nor very high temperatures for this event. It was metamorphosed in the climactic Precambrian event (MP1) and, like the regional migmatites, developed pegmatitic segregations. It was then further deformed by D2, the strongest Caledonian phase, with renewed local segregation caused by the associated high grade metamorphism (e.g. at the Quoich dam - section 17.3.2). D3 structures then formed the present outcrop pattern, the major feature being the Loch Quoich Line, which is entirely irrelevant to the origin of the gneiss.

A synmetamorphic origin for the granitic gneiss would be favoured if it could be derived from Moine metasediments at depth (rather than the lower crust or mantle). In the sense of Chappell & White (1974), the gneiss would appear to be essentially S-type. It has a relatively small volume, with a restricted range of truly granitic bulk compositions with low Na/K ratios, and shows very limited differentiation. It lies in the quartz field, making derivation from an andesitic or more basic magma difficult. It lacks associated basic to intermediate dykes and xenoliths (the dykes at Loch Quoich are part of a regional suite in the Loch Eil Division), but has numerous metasedimentary xenoliths (perhaps locally-derived) and biotitic wisps. It contains two micas (although muscovite is subordinate), is corundum-normative at Quoich, and has accessory garnet, monazite, ilmenite and sulphides, with "sedimentary" zircons containing an old crustal isotopic component. The trace element composition of the Quoich gneiss, including the rare earths, is consistent with derivation from a Moine metasediment (section 16.3.3), or perhaps a Lewisian tonalitic gneiss (cf. the REE pattern in Tarney & Saunders (1979), p. 101), from which the Moines probably derived their REE patterns in any case.

Brook et al. (1976) reported an isochron of 1028±43Ma for the Ardgour gneiss, with an initial Sr^{87}/Sr^{86} of 0.709 - i.e. intermediate between typical S- and I-type values. The Morar Pelite yielded a similar age on what was really an "errorchron", with a similar initial ratio (Brook et al. 1977). The series of pelites studied by Brewer et al. (1979) gave Caledonian whole-rock ages which were attributed to resetting, but the average for each area lay on or close to the Ardgour gneiss isochron. Thus at c. 1000Ma B.P., the average Moine pelite would have had a rather intermediate Sr^{87}/Sr^{86} ratio (0.709), and would be a suitable source rock for derivation, by partial melting, of the granite. The initial ratio of the sediments suggests derivation from either near-contemporaneous volcanogenic material, or Lewisian-type basement plus a short crustal residence prior to 1000Ma.

Pidgeon & Aftalion (1978) found upper and lower intercepts to concordia of 1556 and 574Ma in a U-Pb zircon study of the Ardgour gneiss. However, Aftalion & van Breemen (1980), using Rb/Sr mineral and whole rock ages to independently date the metamorphic episodes, showed that the zircons of both the granitic gneiss and the adjacent metasediments could be explained by a three-stage model, with initial homogenisation at c. 1750Ma, and partial resetting at 1030-1100Ma and c. 490Ma. The Grenville component was dominant in the granitic gneiss, but in the metasediment, Grenville and Caledonian events had caused similar degrees of disturbance. The Moine metasediments therefore contain Laxfordian-type detrital zircons, modified during Grenville and Caledonian events, while the granitic gneiss contains Laxfordian zircons, severely modified in the Grenville, and less so in the Caledonian. The granitic gneiss could have been derived directly by Grenville melting of Lewisian basement, but the "sedimentary" nature of its zircons, and its S-type character, suggest that at least part of its source consisted of Moine metasediments derived by erosion of Lewisian basement. An early pegmatite in the granitic gneiss gave identical results to its host rock, confirming the Precambrian age of these MP1 segregations and of the regional trondhjemitic migmatites.

17.3.2 Caledonian remobilisation of the granitic gneiss

A prominent feature of the granitic gneiss in the Quoich dam section is the development of shear zones and conjugate folds, associated with localised segregation or partial melting (Chapter 11). Some conjugate structures seem to show a large volume increase, with concentration of pegmatitic material in the cores, while the shear zones are coarsely recrystallised, but may have suffered a volume decrease (Plate 11.2(c)). In some cases, pegmatites extend from them and intrude the nearby granitic gneiss or basic dykes (Plate 11.1(a)). In the vicinity of the shear zones and pegmatites, the normal equigranular texture of the granitic gneiss is severely modified.

Near a shear zone margin (Plate 17.3(c)), feldspars are fractured and quartz recrystallised, and grain boundaries develop a thin coating of fine-grained quartz and alkali feldspar. This tends to be thicker (up to 50 microns) on boundaries lying at a high angle to the foliation. The adjacent plagioclase may show very slight reverse zoning. This can be interpreted as incipient melting at grain boundaries, with preferential segregation of melt on those which have the lowest resolved normal stress (compression, indicated by the sense of shear zone movement, being at a high angle to banding). In the centre of a shear zone (Plates 17.3(d), 17.4(a)), there is a large range of grain sizes, including "old" 1mm plagioclases which appear to be corroded and certainly show marginal reverse zoning. There is a much greater proportion of the quartz/alkali feldspar material, some of which is quite coarse-grained, and contains rotated fragments of relict plagioclase (Plate 17.3(d)). The probable melt film is no longer restricted to a particular orientation of grain boundary.

The shear zones therefore appear to be loci for a greater degree of partial melting than in the surrounding granite gneiss, perhaps due to concentration of volatiles there, either penetrating from the adjacent pelite, or drawn from the surrounding granitic gneiss by dilatancy pumping (see below). The concentration of these phenomena in the eastern part of the granitic gneiss (the top, prior to D3), adjacent to a thick pelite, suggests that ingress of water to the granitic gneiss was important.

These features can be compared with the experiments of van der Molen & Paterson (1979), who deformed a partially melted sample of Delegate Aplite, a rock very similar in composition and grain size to the Quoich gneiss. They found that, under hydrostatic conditions, no more than about 10% melt formed (regardless of the water content), at which point 80-90% of grain boundaries had a melt film, and the rock was mechanically a granular aggregate. However, if the rock was deformed, melting approached the theoretical maximum for a given H_2O content - i.e. strain created interstices into which melt could migrate. At low strains, melt was concentrated in fractures and grain boundaries parallel to the maximum principal stress. At higher strains, shear zones developed by cataclastic flow, lubricated by the melt film, which formed up to 20% by volume of the zone.

A theoretical analysis by van der Molen & Paterson (op.cit.) indicated that, beyond 25-40% melt (depending on grain-size distribution), the strength decreases sharply by several orders of magnitude, and even with 25% melt, the rock is perhaps 100 times less strong than with 10% melt. A granular flow mechanism for the deformation implies an initial volume increase as strain produces less efficient packing of the grains, which gives rise to dilatancy pumping (the drawing-in of fluid from the less deformed rock, into the shear zone). Their experiments were carried out at 3kb, 800°C, with strain rates of about 10^{-5} s^{-1} - the last two values probably significantly greater than those which obtained in the deforming granitic gneiss.

The shear zones at Loch Quoich, with their high melt fraction, were therefore much weaker than the surrounding gneiss, and once initiated, would have taken up most of the bulk strain.

From section 16.3.3, a shear zone sample (Q13b) actually has a restitic composition. This must mean that, although a large proportion of melt remains in the rock, considerably more has been expelled - e.g. if the melt had subequal quartz, K-feldspar and albite, the change in normative plagioclase from An_{11}(granitic gneiss) to An_{16} (shear zone) implies loss of perhaps 30% melt by volume. This rock would therefore represent the residue, after about 50-60% partial melting, plus about half of the melt, which was trapped between the restite grains.

In van der Molen & Paterson's (1979) experiments the shear zones concentrated melt from the surrounding rock, since local melting gave rise to less than the critical 25-30% melt. Since the felsic components of the near-cotectic granitic gneiss could be almost completely melted at only slightly above the solidus temperature (P_{H_2O} = P_{load}), the degree of melting should be controlled by the a vailability of H_2O. If the shear zone drew in H_2O (and possibly some melt) from the country rock, or acted as a channel for aqueous fluids, the melt fraction would rapidly increase to >30%, the strength would fall to a fraction of a bar, and the resolved normal stress on the shear zone would result in expulsion (filter pressing) of the excess melt to form discrete pegmatites. The equilibrium melt content of the shear zone would be determined by the balance between the strength of the crystal mush and the resolved normal stress on the shear zone boundaries, at the transition from dilatancy pumping to filter pressing.

Pegmatitic or granitoid areas in conjugate fold cores, and diffuse pegmatitic patches within the eastern part of the granitic gneiss, show similar petrographic features to the shear zones (Plate 17.4(b), (c)). The main differences are the relative lack of strain, the greater proportion of the quartz/albite/K-feldspar component, and the very large grain size (several centimetres) of some of the alkali feldspars. Swap textures, like those in Q22a, are also common. These too appear to be melt/restite or melt/palaeosome mixtures, some presumably arising by collection of melt expelled from shear zones, others produced by widespread minor melting in the granitic gneiss, and concentration of melt in favourable sites.

One of the large, late granitoid patches was analysed (Q1 - Plate 17.4(b)). It has a very albitic composition, and is clearly a "leucosome" relative to the granitic gneiss (section 16.3.3), but is too K-feldspar - poor to have a low melting temperature (perhaps a sampling problem), and also appears to be in subsolidus trace element equilibrium with the granitic gneiss. It is possible that some of these bodies were produced by subsolidus segregation, continuing after melting had ceased, perhaps promoted by Na-rich fluids released during crystallisation. Alternatively, large granitoid bodies may have crystallised more slowly, with release of large volumes of fluid which could not readily escape when deformation had ceased, and caused subsolidus re-equilibration of the gneiss and the crystallised melt. Restite material, like Q13b, would have had much of its vapour phase removed with the melt, and so might fail to re-equilibrate during cooling.

17.4 Late migmatisation

The term "late migmatisation" has been used rather broadly to describe a range of phenomena which are relatively late in the structural sequence. In areas near the Sgurr Beag Slide, this means post-sliding or (rarely) syn-sliding. Elsewhere, it refers to events later than the local D2, which probably correlates broadly with sliding (section 14.3). In practice, this implies that the late migmatisation is Caledonian, as distinct from the Precambrian regional migmatisation (section 14.4). It can be divided into the products of local segregation, and of Caledonian pegmatite emplacement.

17.4.1 Local segregation

This occurs in the more easterly areas, where Caledonian metamorphic grade was relatively high. Examples were described from Kinlochourn, Acharacle, Sguman Coinntich and Loch Quoich (sections 5.4.2, 6.4.2, 8.4.2, 11.3.2). Generally, migmatisation was triggered by deformation, and is particularly marked in shear zones and highly-deformed fold limbs, but occasionally took place along fold axial planes. Examples from the various areas can be seen in Plates 8.1(f), 8.4(d), 11.1(d)-(f), 17.4(d). Occasionally, similar phenomena appear to be independent of structure (Plate 6.4(a),(h)), although they may have been influenced by fluid movement along fractures.

Late migmatisation tends to be restricted to psammites and semipelites which were not highly-migmatised by the Precambrian event, and the typical pelitic gneiss only displays it when the texture had previously been broken down by very strong deformation, or in areas of unusually high Caledonian grade (e.g. Loch Quoich). Following the arguments of section 17.1, the regional migmatites have achieved a stable texture, while the more non-migmatitic lithologies still have mixed quartz/plagioclase/mica assemblages suitable for migmatisation. This is particularly true where Morar Division psammites (not previously migmatised) were affected by relatively high grade Caledonian metamorphism (e.g. at Sguman Coinntich and Kinlochourn). The restriction of migmatisation to zones of relatively high strain presumably results from the dry, already-metamorphosed nature of the psammites. Grain-size reduction, and increase in fluid content due to dilatancy pumping (cf. section 17.3.2) or actual fluid flow through shear zones and fold limbs, would enhance any diffusion-related process.

The local derivation of these veins is confirmed by their often-blind nature, the presence of a mafic selvedge, and their compositional control by the host rock. K-feldspar is only present if the host rock contains it, and epidote is common in veins derived from psammites. The plagioclase composition in the vein corresponds to that in the host rock (cf. K15a and K16 in Appendix 2), although the grain size is often much larger (e.g. 1cm <u>vs</u>. 1-2mm). The apparent equilibrium between host and vein, and the widespread distribution of these migmatites in areas of only moderately high grade, suggest a subsolidus mechanism of formation.

However, segregations are restricted by metamorphic grade - they occur east of a line, essentially independent of structure, accross which the grain size of rocks deformed in D2 and D3 increases markedly, and strain features are lost. This is particularly well displayed in the Acharacle area (section 6.4.2). The segregations discussed in sections 17.2 and 17.3.2 are of this age, but developed where Caledonian temperatures were unusually high and permitted partial melting; the Lochan Coire Shubh complex (see below) formed by a combination of pegmatite intrusion and local partial melting.

17.4.2 Caledonian pegmatites

For the most part, these tend to be distinctly later than the segregations discussed above, and are virtually undeformed. They are distinguished by their coarse grain-size, the common presence of K-feldspar and albite, their sharply cross-cutting and intrusive relationships, their independence of host-rock composition, and the lack of a mafic selvedge. Although they generally occur in the zone of relatively high Caledonian grade discussed above, they are geographically restricted within this zone, and the veins within each area have a consistent character, which is not obviously related to the local rock types or metamorphic grade. Examples were discussed in sections 5.4.2, 6.4.2, 8.4.2, 9.4.4, 10.3, and 11.3.2 - cf. Plates 5.1(g),(h); 6.1(c), 9.1(h), 9.3(d), 10.1(d), 11.1(h), 17.3(b). Some look like subsolidus veins (e.g. Plate 5.1(g)), but others were probably volatile-rich melts derived from intrusions or partial melts at greater depth.

On Ben Klibreck, these veins form a late-tectonic suite ranging from hornblende-granodiorite to albite-bearing pegmatites, and are probably related to some nearby Newer Granite (section 9.4.4). Pegmatites are relatively rare at Acharacle and Sguman Coinntich, but appear to be similar to the Kinlochourn variety. At Loch Quoich, the veins are very acid (K-feldspar/albite - rich), and the latest ones have two micas, suggesting a deeper Moine source, or at least some metasedimentary influence.

The Lochan Coire Shubh complex may be a good model for the source of many of the Caledonian pegmatites in the Glenfinnan Division (Chapter 10). There, intrusions of pegmatites from greater depth was accompanied by local partial melting of K-feldspar - bearing metasediments. The sequence of events, defined by cross-cutting relationships and the degree of $D3_1$ strain, is listed below:

Intrusion of pegmatites (carrying micaceous restite) into the $F3_1$ fold core (Plate 10.2(c)).
Partial melting of metasediments in highly-strained $F3_1$ fold limbs with production of local restite, and continued intrusion of pegmatites.
Schollen migmatisation, again involving both local and introduced material.
Intrusion of a suite of dykes, axial planer to the $F3_1$ fold (Plate 10.3(a)).
The climax of local partial melting, expulsion of some of the melt leaving areas of restite (Plate 10.3(b)), and agmatisation of psammites by locally-derived quartz-rich aplites (Plate 10.3(c)).
Intrusion of a regional dyke swarm (Plate 10.3(d)); these pegmatites contain restite patches, but have sharply cross-cutting contacts.

In the early stages of development of the complex, pegmatites, perhaps produced at only slightly greater depth, were intruded. Later, intrusion continued, but as isotherms rose (?) local partial melting also occurred; towards the end of this phase, expulsion of locally-derived melt took place. Finally, melting ceased at the present level of exposure, but continued at depth to produce the latest dykes. The more distal dykes are similar to those at Kinlochourn (2km away), and the chemistry of one of these (LH23 - section 16.3.4) is consistent with derivation by partial melting of metasediment.

This history suggests that the present level of exposure intersects the top of a local, late Caledonian thermal dome in which melting took place; at depth, melting would have been more severe and longer-lasting, and could have given rise to far-travelled pegmatites in the surrounding Moines.

CHAPTER 18

Summary of conclusions

18.1 Structure

The subdivision of the Northern Highland Moines into Morar, Glenfinnan and Loch Eil divisions (Johnstone et al. 1969, Johnstone 1975) is accepted. The Morar and Glenfinnan divisions are separated by the Sgurr Beag Slide (section 14.2), with the higher-grade Glenfinnan Division being thrust over the Morar Division. No comparable structural or metamorphic break is present between the Glenfinnan and Loch Eil divisions (sections 11.4, 12.3), and a normal stratigraphic passage seems likely (the Loch Eil Division being younger). The Loch Quoich Line (Clifford 1958a, Roberts & Harris in press) is related to the last regional folding event, and defines the boundary between a western area where these upright folds are tight and steeply-plunging, and an eastern area (mainly Loch Eil Division) where they are open and gently-plunging.

The Sgurr Beag Slide is a major tectonic break, separating rocks of contrasting metamorphic grade and with completely different stratigraphies; over much of its length, it forms the western boundary to regional migmatisation (section 14.2). Sliding has imposed an intense LS fabric on the rocks on either side, which is consistent with westerly-directed simple shear. Displacements of at least 40-80km by westerly overthrusting are indicated (sections 7.4, 14.2), and progressively younger Morar Division rocks underlie the slide in the west. Isoclinal folds in the Glenfinnan Division are probably associated with the slide (rather than the slide being subordinate to the folds), but contemporaneous deformation in the Morar Division is restricted to bedding-parallel shear zones or grain-size reduction zones. A west-to-east sequence can be observed in the deformation associated with the slide, from a shear zone cutting essentially undeformed Morar and Glenfinnan division rocks, to a deformation front with isoclinal folding above the slide, restricted shear zones and subsidiary slides beneath it. This was interpreted in section 14.2 as a sequence from shallower to deeper levels of the slide. The Knoydart Slide, and those slides (associated with Lewisian inliers) which underlie the main slide on Ben Klibreck and Sguman Coinntich, were considered to bound duplexes produced beneath and ahead of the nappe carrying the Glenfinnan Division.

The intense platy fabric was used as a structural marker for correlation along and across the Sgurr Beag Slide. It was considered in section 14.3 that the most complete structural sequence is preserved in the western Morar Division, Knoydart Pelite and Mull Glenfinnan Division. There, an early bedding-parallel fabric (axial planar to rare isoclinal folds) was crenulated to form the dominant (often penetrative) S2 schistosity. Garnet at Mallaig is MP1-MS2, but in Knoydart MP1, MS2 and widespread MP2 garnet was recorded, and in Mull, garnet is at least MP1 and some may be MP2. Quartz ($\pm$minor feldspar) veins of MP1 age are common in all these areas, but migmatisation (segregation of feldspar-rich quartzofeldspathic lits) is MS2-MP2 in Knoydart and Mull. The slide fabric, and the analogous shear zones and grain-size reduction zones at Mallaig, are of D3 age in this sequence, and the Morar Antiform and Assapol Synform (which folds the slide on Mull) at least D4. In most other areas, the slide was assigned to local D2, because only one earlier fabric could be distinguished. However, it was argued in section 14.3 that the weak S1 fabric detected in the west could not be recognised in these higher-strain areas, and that this eastern S1 corresponds to a composite S1-S2 of the western sequence. This also applies to the main outcrop of the Glenfinnan Division, where isoclinal (local) F2 folds are refolded by tight, upright F3 folds, analogous to those which fold the slide. These folds have curvilinear axes and upright, N-S or NNE-SSW axial planes in the southern Moines, but in the NW Highlands they are replaced by reclined sheath folds with a gently SE-plunging extension lineation, comparable to D2 of the Moine Thrust Belt. On Ben Klibreck, the local sequence had sliding as $D1_b$, "Sutherland D2" (e.g. Soper & Brown 1971) as $D2_b$, but there is evidence for a pre-$D1_b$ fabric in both Morar and Glenfinnan divisions, and regional migmatisation in the latter is pre-$D1_b$.

Thus local D2 is successively younger in the sequence SW Inverness-shire - E Inverness-shire - Sutherland.. This reflects increasing strain associated with the later events, and not a real difference in the age of the rocks. Once a particular tectonic event has produced isoclinal sheath folds with strong LS fabrics, it is extremely unlikely that earlier events, however pervasive originally, can be separated - i.e. local D2 will in general be the last event associated with high ductile strains, and may be D3 or D4 in an absolute sequence. In the case of the main Glenfinnan Division outcrop, it is the "Sgurr Beag Slide" event which, in producing intense deformation in the rocks to the east of the slide, has made earlier events difficult to recognise. In W Sutherland (which lay west of this deformation front) the rocks have been severely affected by later deformation related to movement on the Moine mylonites. In SW Inverness-shire, both these events gave rise to high strains only locally, and the full pre-sliding structural sequence can be recognised.

In section 14.4, this structural sequence was used to interpret published radiometric dates. It was concluded that all three divisions of the Moines have suffered Precambrian metamorphism and deformation (corresponding to D1 and D2 of Mallaig), and that this probably took place during a Grenville (c. 1000Ma) rather than a Morarian (c. 750Ma) event. The lit-par-lit migmatites of the Glenfinnan Division and parts of the Morar Division formed during this event, but the cross-cutting Morarian pegmatites, which have yielded many 700-800Ma dates, may represent a distinct intrusive event not associated with regional deformation. Alternatively, slow cooling until post-Grenville uplift at c. 750Ma may be responsible for these dates. The Sgurr Beag Slide and local D2 of the main Glenfinnan Division outcrop (D3 of Mallaig) are considered to be early Caledonian (c. 460Ma?), while the upright (local) D3 folds of this area, and D2 of Sutherland, are synchronous with early movement in the Moine Thrust Belt ("late Caledonian"), perhaps at c. 440Ma.

Thus in the SW Morar Division, the dominant metamorphic fabric is Precambrian, but the main map-scale structures, such as the Morar Antiform, are Caledonian. In the Glenfinnan Division, migmatitic banding and the metamorphic peak are Precambrian, but the dominant fabric is Caledonian, and was originally recumbent, while the major upright structures (which define the present dip of bedding) and the Loch Quoich Line are late Caledonian. In W Sutherland, the early metamorphic and migmatitic fabric is Precambrian, as are at least some garnets, but the major structures and the widespread LS fabric are late Caledonian, although high early Caledonian strains affect some slide zones in the eastern Morar Division.

18.2 Metamorphism

Caledonian metamorphic grade (syn- to post-sliding) increases fairly uniformly to the east or southeast, from biotite grade in western Morar to perhaps garnet or staurolite grade at the main outcrop of the Sgurr Beag Slide, with higher temperatures occurring locally (e.g. at Loch Quoich, where partial melting took place - sections 17.2, 17.3.2), or being associated with granite or pegmatite intrusion (e.g. Ben Klibreck, Lochan Coire Shubh - Chapters 9, 10).

In the Morar Division, Precambrian metamorphic grade rose from garnet in the west to perhaps kyanite in the easternmost outcrops exposed, although grade based on mineral assemblages is difficult to determine because typical Moine pelites will not produce aluminosilicates until very high temperatures have been reached (section 16.1). For this reason, an extensive program of geothermometry and geobarometry was undertaken (Chapter 15), using Thompson's (1976) garnet-biotite thermometer and a variety of barometers. The garnets used were texturally early (Precambrian); the preservation of zoning, and the consistent jump in temperature across the slide, suggests that reasonable estimates were obtained for the Precambrian metamorphism, despite Caledonian overprinting. Although MacQueen & Powell (1977) have suggested that some chemically-unzoned garnets were produced by Caledonian homogenisation, the detailed distribution of zoned and unzoned garnets was considered in section 15.1.2 to favour other explanations. In the Morar Division (section 15.2.2), results of c. 520°C, 6.5kb were obtained for MS2, with MP1 possibly nearer 550°C, and MP2 near 460°C, 5.8kb. The lowest estimate (470°C, 4-4.5kb) was obtained from western Ardnamurchan (possibly biotite grade - Butler 1965), while results for the eastern Morar Division clustered around 600°C, 6kb. The structurally-deepest sample (from Ben Klibreck) gave 615°C, 7kb, although the significance of this may be questioned, since the resolution of the barometers is only of the order of 1kb. Glenfinnan Division samples from immediately above the Sgurr Beag Slide gave results of 630-640°C at 6-6.5kb (section 15.2.3), while structurally-higher (Loch Quoich) and more westerly (Sguman Coinntich, Ross of Mull, Knoydart Pelite) samples gave lower temperatures (580-600°C at 5.5-7kb).

18.3 Migmatisation

Migmatisation in the Moines embraces many phenomena (e.g. section 6.4, Chapter 17), and much of the conflicting literature has arisen from a failure to differentiate these. Early migmatites (pre-Sgurr Beag Slide) are essentially lit-par-lit pelitic gneisses (the "Permeation Gneiss" of the Geological Survey) with K-feldspar - free leucosomes; these have been called the "regional trondhjemitic migmatites", and with the garnets discussed above are relics from a Precambrian (Grenville?) metamorphism. Their mineralogy, lack of feldspar differentiation and the sub-650°C temperature estimates are inconsistent with a partial melting origin (cf. section 15.3.1 for model melting reactions and section 17.1 for petrography and discussion of subsolidus segregation models). Migmatisation is controlled by chemistry and/or lithology - plagioclase-rich rocks, especially the Garnetiferous Pelite, are the most readily migmatised, while micas inhibit migmatisation (section 16.2). High K_2O does not favour migmatisation, and Na/Na+Ca has no effect, providing further arguments against a partial melting mechanism.

Later migmatites include a variety of Caledonian pegmatites, some intrusive (e.g. at Lochan Coire Shubh), others locally-derived (e.g. at Acharacle and Sguman Coinntich). On Ben Klibreck, Brown (1967) and Soper & Brown (1971) have indiscriminately sampled regional trondhjemitic migmatites (restricted to the Glenfinnan Division), more widespread Caledonian pegmatites and a series of calc-alkaline granodiorites and granites which are probably related to the Vagastie Suite of Read (1931); if these are treated separately, there is no need to invoke metasomatism, although the intrusive members naturally changed the bulk composition on the scale of an outcrop.

The West Highland granitic gneiss (section 17.3) is a deformed and metamorphosed granite, and was intruded into both Glenfinnan and Loch Eil divisions prior to the main Precambrian deformation, probably as a series of sills. Previous interpretations as a migmatitic or metasomatic product (e.g. Dalziel 1966) resulted from a failure to distinguish between the gneiss itself, segregations produced from it during later metamorphism, and much later K-feldspar - bearing aplites and pegmatites, and to recognise transposition of lits and segregations into strong tectonic fabrics. It is a typical S-type granite (sensu Chappell & White 1974), and could have been produced by Grenville partial melting of Moine-like rocks at greater depth (section 17.3.1 (iii)).

Local Caledonian partial melting has occurred in the vicinity of the Loch Quoich spillway, where unusual K-feldspar - bearing migmatites were produced from Loch Eil Division pelites and the granitic gneiss (sections 17.2, 17.3.2), and near Lochan Coire Shubh, where pegmatites (possibly partial melts of Moine metasediments - sections 16.3.4, 17.4.2) intrude rocks which themselves appear to have undergone partial melting.

18.4 Regional models

In this final section, some speculative models will be derived which attempt to explain the broad distribution of lithology and metamorphic grade in those parts of the Moines which were investigated in the course of this thesis. In Chapter 14, it was concluded that the Sgurr Beag Slide and associated structures were best interpreted as early Caledonian thrusts disrupting a Precambrian metamorphic complex, with only detail modifications resulting from later folding.

The models which will be developed follow Dahlstrom's (1970) rules for thrusting - i.e. thrusts (comprising a series of ramps and flats) climb in the transport direction, placing deeper rocks over shallower, and propagate in "piggy-back" fashion towards the foreland (cf. also Butler 1982b and Boyer & Elliott 1982). This neglects the possible fold-nappe character of the deeper Glenfinnan Division (which only applies to the hanging-wall) and assumes that structural/stratigraphic units within the Morar Division were broadly horizontal at the end of the Precambrian event (perhaps not unrealistic, given the fold-nappe character generally inferred for the Precambrian structures - e.g. Powell 1974, fig. 5). The effect of post-slide folding can most conveniently be represented by manipulating the "erosion surface" on the final sections - i.e. the thrusts will be shown as essentially flat-lying, and the "erosion surface" will have the inverse geometry to the post-slide structures (neglecting topography). The sections are developments of those presented in Chapter 14, with additional constraints on the initial positions of hanging-wall units being provided by the temperature estimates of Chapter 15.

Some general statements can be made, based on the regional geology of the Moines. The relative uniformity in lithology and metamorphic grade shown by the Glenfinnan Division overlying the Sgurr Beag Slide indicates that the slide was sub-parallel to bedding and metamorphic zones, i.e. that the Glenfinnan Division originally overlay an extensive flat - perhaps following the Moine/Lewisian boundary. From Kinlochourn to Lochailort, younger units of the Morar Division underlie the slide, so there must be a ramp or series of ramps in this area (cf. section 14.2) - thus progressively older and higher-grade Morar Division rocks underlie the slide towards the east, and the metamorphic contrast with the Glenfinnan Division decreases. Figure 18.1(a) shows a sketch of this situation, with the Knoydart Pelite (1) forming a duplex derived from the Kinlochourn ramp (2,5), and the Ben Klibreck Morar Division (3) transported from a more easterly ramp (4), perhaps that on which the slide climbed from the top of the Lewisian to the top of the Lower Morar Psammite (cf. Figs. 14.4, 14.5, 14.6). The Glenfinnan Division would be somewhat thinner than the height of the ramp - a few kilometres - which is not unreasonable from inspection of I.G.S. one-inch maps. Given a fairly horizontal basement-cover interface, the broad distribution of Morar, Glenfinnan and Loch Eil divisions could have been defined by this structure (see inset, Fig. 18.1(a)), although it is not sufficient in itself to explain the steep belt/flat belt contrast, the zone of ramps climbing by only a few kilometres in 10-15km, i.e. at an average dip of about 10°. However, folds associated with the ramps and perhaps related duplexes may have been tightened or served as nuclei during later compression and/or simple shear related to movement on the Moine Thrust.

Restoration of Fig. 18.1 (a) begins by replacing the westernmost (Knoydart) duplex (1) in its proper stratigraphic position - its 580°C temperature implying one intermediate between Inverie and Kinlochourn (Fig. 18.1(b) - cf. Fig. 14.5(b)). The Ben Klibreck duplex (3) is then restored, as in Fig. 18.1(c) - cf. Figs. 14.4(b), 14.6. This represents the position of the Sgurr Beag Slide before duplexes were removed from the Morar Division of the footwall. The Glenfinnan Division must then be moved further east, so that the 600°C Mull Glenfinnan Division (6) aligns with the c. 600°C Morar Division of Acharacle/Kinlochourn (7). Fig. 18.2(a) shows such a restoration with suggested isotherms and broad stratigraphic units added. This assumes that the Glenfinnan Division is the distal equivalent of the Lower Morar Psammite and the Morar Schist, and the Loch Eil division of the Upper Morar Psammite (the overall pelite/psammite ratio increasing towards the east). 10-15km west of Ben Klibreck, the Loch Meadie schists of Read (1931)- 8 - are kyanite-bearing and rest on the Meadie Slide (Soper & Brown 1971, Soper & Barber 1982, V.E. Moorhouse verb. comm.). Along with the Morar Division lying west of the zone of abundant Lewisian inliers (cf. Chapter 9), they could form a duplex analogous to that in Knoydart; the presence of kyanite suggests affinities with the westernmost exposed Glenfinnan Division on Mull, so they have been placed between the Knoydart Pelite (1) and the Mull Glenfinnan Division (6). Clearly other models are possible - e.g. the Glenfinnan Division could be much further-travelled, and unconnected with the Morar Division metamorphic complex, but this is the simplest model which explains the available data, and could be regarded as an intermediate stage in a model with very far-travelled Glenfinnan Division. Note that this model reduces the importance of the Morar/Glenfinnan division distinction - e.g. it implies that psammites (3) which are usually correlated with the Morar Division (on Ben Klibreck) were overlain by rocks equivalent to the Ross of Mull Glenfinnan Division. This may not be unreasonable - the psammites have no diagnostic features, while the Mull Glenfinnan Division (especially the striped pelites and psammites) has some affinities with the Morar Division. The distinctive garnetiferous ampibolites of Mull and the southern Glenfinnan Division are absent on Ben Klibreck, but this could be a north-south rather than an east-west feature. This model therefore has some affinities with that in Fig. 14.6(d), as well as that in Fig. 14.6(c).

Fig. 18.2(b)-(d) shows a model for the structural development of the area west of Ben Klibreck. Firstly, higher-grade Glenfinnan Division (9) is superimposed on the Ben Klibreck Morar Division (3); then this Morar Division (3) is transported as a duplex onto the Loch Meadie schists (8). These are finally emplaced onto the low-grade Morar Division (10) overlying the Moine Thrust. This model may also explain the high-grade character of the slide on Ben Klibreck (cf. section 14.2) - it is a "fossil slide" which has been transported west by movement on deeper structures. On Carn Chuinneag, the Morar Division appears to have reached only garnet grade during the Precambrian event (Wilson & Shepherd 1979) - this could be produced from Fig. 18.2(a) by a model having no duplexes (Fig. 8.2(e)), so that high-grade Glenfinnan Division rocks (9) rest directly on the sole thrust (11). Duplexes beneath the Moine Thrust often have limited lateral extents (e.g. in Assynt, McClay & Coward 1981), and such a wedgingout is not inconsistent with regional maps of the area between Ben Klibreck and Carn Chuinneag.

Fig. 18.3 contains an analogous series of sections for the southern Moines. The Loch Meadie schists (8) are removed with the Glenfinnan Division and lie west of the Mull rocks (6 - Fig. 18.3(a), (b)). To represent the Mallaig-Kinlochourn section, the Knoydart Pelite (1) is then transported as a duplex (Fig. 18.3(c)), and Glenfinnan Division rocks (9) overlie the high-grade Lower Morar Psammite at Kinlochourn (7). The Mull to Loch Eilt section is represented by Fig. 18.3(d), (e) - the high-grade Morar Division (12) has not formed a duplex, and remains in the footwall; high-grade Morar Division (12) underlies the Glenfinnan Division (9) at Acharacle/Salen and the Mull Glenfinnan Division (6) overlies the very low-grade Morar Division of the extreme west (13). This model assumes that the rocks in the Ardnamurchan Antiform (the en echelon continuation of the Morar Antiform) are in stratigraphic continuity. This may not be so - the I.G.S. one-inch map cannot readily be explained in terms of the Morar stratigraphy, there are several high-strain zones within the Morar Division (O'Brien 1981), and there is a sharp P,T contrast across the antiform (section 15.2.2(iii)). In that case, the Salen-Loch Eilt rocks (12) may have been transported in a further duplex (analogous to that containing the Knoydart Pelite) to lie partly on the Knoydart Pelite (cf. Fig. 14.5(a)) and partly on the unmoved Morar Division (in practice, the Knoydart duplex may wedge out to the south, and the Loch Eilt duplex to the north, so that only a slight overlap is present).

The only area which cannot readily be explained is Sguman Coinntich (Chapter 8), where the Glenfinnan Division apparently occurs in the core of an antiform. If, on rare occasions, a deeper thrust was allowed to cut through an older, shallower thrust, one possibility would be to allow such a thrust (west of those shown in Figs. 18.2, 18.3) to emplace low-grade Morar Division rocks on Mull-type Glenfinnan Division.

In conclusion, the regional geology of the Moines can be explained by the disruption of an earlier stratigraphic/metamorphic complex by early Caledonian thrusts, although much more information would be required to draw properly-balanced sections, and the presence of pre- and post-thrusting deformation events is bound to produce exceptions to the overall pattern.

REFERENCES

AFTALION, M. & VAN BREEMEN, O. 1980. U-Pb zircon, monazite and Rb-Sr whole-rock systematics of granitic gneiss and psammitic to semipelitic host gneiss from Glenfinnan, northwestern Scotland. Contrib. Mineral. Petrol. 72, 87-98.

ALLMENDINGER, R. W. 1981. Structural geometry of Meade thrust plate in northern Blackfoot Mountains, Southeastern Idaho. Bull. Am. Assoc. Petrol. Geol. 65, 509-25.

ASHWORTH, J. R. & CHINNER, G. A. 1978. Coexisting garnet and cordierite in migmatites from the Scottish Caledonides. Contrib. Mineral. Petrol. 65, 379-94.

ATHERTON, M. P. & TARNEY, J. (eds.) 1979. Origin of granite batholiths: geochemical evidence. Shiva Publishing, Kent.

BAILEY, E. B. 1910. Recumbent folds in the schists of the Scottish Highlands. Q. J. geol. Soc. London. 66, 586-608.

BAILEY, E. B. & ANDERSON, E. M. 1925. The geology of Staffa, Iona and western Mull. Mem. geol. Surv. U. K.

BAIRD, A. W. 1982. The Sgurr Beag Slide within Moine rocks at Loch Eilt, Inverness-shire. J. geol. Soc. London. 139, 647-53.

BAMFORD, D., NUNN, K., PRODEHL, C. & JACOB, B. 1977. LISPB-III. Upper crustal structure of Northern Britain. J. geol. Soc. London. 133, 481-8.

BAMFORD, D., NUNN, K., PRODEHL, C. & JACOB, B. 1978. LISPB-IV. Crustal structure of Northern Britain. Geophys. J. R. astron. Soc. 54, 43-60.

BARBER, A. J. 1965. The history of the Moine thrust zone, Lochcarron and Lochalsh, Scotland. Proc. Geol. Assoc. London. 76, 215-42.

BARBER, A. J. & MAY, F. 1976. The history of the Western Lewisian in the Glenelg Inlier, Lochalsh, Northern Highlands. Scott. J. Geol. 12, 35-50.

BECKINSDALE, R. D. & OBRADOVITCH, J. D. 1973. K-Ar ages for minerals from the Ross of Mull, Argyllshire, Scotland. Scott. J. Geol. 9, 147-56.

BOWES, D. R. 1968. The absolute time-scale and the subdivision of Precambrian rocks in Scotland. Geol. För. Stöckh. Förh. 90, 175-88.

BOWES, D. R. 1980. The absolute time-scale and the subdivision of Precambrian rocks in North-western Britain. In MITROFANOV, F. P. (ed.) Principles and criteria of subdivision of Precambrian in mobile zones. Leningrad: Nauka.

BOWES, D. R. & LEAKE, B. E. (eds.) 1978. Crustal evolution in northwestern Britain and adjacent regions. Geol. J. Spec. Issue. 10.

BOYER, S. E. & ELLIOTT, D. 1982. Thrust systems. Bull. Am. Assoc. Petrol. Geol. 66, 1196-1230.

BRADBURY, H. J. 1979. Migmatisation, deformation and porphyroblast growth in the Dalradian of Tayside, Scotland. In HARRIS, A. L. et al. (1979), pp. 351-6.

BRADBURY, H. J., SMITH, R. A. & HARRIS, A. L. 1976. "Older" granites as time-markers in Dalradian evolution. J. geol. Soc. London. 132, 677-84.

BRADBURY, H. J., HARRIS, A. L. & SMITH, R. A. 1979. Geometry and emplacement of nappes in the Central Scottish Highlands. In HARRIS, A. L. et al. (1979), pp. 213-20.

BREWER, M. S., BROOK, M. & POWELL, D. 1979. Dating of the tectonometamorphic history of the southwestern Moine, Scotland. In HARRIS, A. L. et al. (1979), pp. 129-37.

BROOK, M., POWELL, D. & BREWER, M. S. 1976. Grenville age for rocks in the Moine of north-western Scotland. Nature, London. 260, 515-7.

DEARNLEY, R. 1967. Metamorphism of minor intrusions associated with the Newer Granites of the Western Highlands of Scotland. Scott. J. Geol. 3, 449-57.

DREVER, H. E. 1940. The geology of Ardgour, Argyllshire. Trans. R. Soc. Edinburgh. 60, 141-70.

DUNHAM, A. C. & WILKINSON, F. C. F. 1978. Accuracy, precision and detection limits of energy dispersive electron microprobe analyses of silicates. X-Ray Spectrometry. 7, 50-56.

DUNNING, F. W. 1972. Dating events in the metamorphic Caledonides: impressions of the symposium held in Edinburgh, September 1971. Scott. J. Geol. 8, 179-92.

ELLIOTT, D & JOHNSON, M. R. W. 1980. The structural evolution of the northern part of the Moine thrust zone. R. Soc. Edinburgh Trans. Earth Sci. 71, 69-96.

ESCHER, A. & WATTERSON, J. 1974. Stretching fabrics, folds and crustal shortening. Tectonophysics. 22, 223-31.

EUGSTER, H. P., ALBEE, A. L., BENCE, A. E., THOMPSON, J. B. & WALDBAUM, D. R. 1972. The two-phase region and excess mixing properties of paragonite-muscovite crystalline solutions. J. Petrol. 13, 147-79.

FERGUSSON, I. W. 1978. Structural age of the Vagastie Bridge granite. Scott. J. Geol. 14, 89-92.

FERRY, J. M. & SPEAR, F. S. 1978. Experimental calibration of the partitioning of Fe and Mg between biotite and garnet. Contrib. Mineral. Petrol. 66, 113-7.

FETTES, D. J. 1979. A metamorphic map of the British and Irish Caledonides. In HARRIS, A. L. et al. (1979), pp. 307-21.

FETTES, D. J. & McDONALD, R. 1978. Glen Garry vein complex. Scott. J. Geol. 14, 335-58.

FITCH, F. J., MILLER, J. A. & MITCHELL, J. G. 1969. A new approach to radio-isotopic dating in orogenic belts. In KENT, P. E. et al. (1969), pp. 157-95.

FLEUTY, M. J. 1961. The three fold-systems in the metamorphic rocks of Upper Glen Orrin, Ross-shire and Inverness-shire. Q. J. geol. Soc. London. 117, 447-76.

FLEUTY, M. J. 1964a. The description of folds. Proc. Geol. Assoc. London. 75, 461-89.

FLEUTY, M. J. 1964b. Tectonic slides. Geol. Mag. 100, 244-54.

FLEUTY, M. J. 1974. The Bheinn Bhreac Fold, Southern Ross-shire. Scott. J. Geol. 10, 229-35.

GANGULY, J. 1979. Garnet and clinopyroxene solid solutions, and geothermometry based on Fe-Mg distribution coefficient. Geochim. cosmochim. Acta. 43, 1021-9.

GANGULY, J. & KENNEDY, G. C. 1974. The energetics of natural garnet solid solution. I. Mixing of the aluminosilicate end members. Contrib. Mineral. Petrol. 48, 137-48.

GHENT, E. D., ROBBINS, D. B. & STOUT, M. Z. 1979. Geothermometry, geobarometry and fluid compositions of metamorphosed calc-silicates and pelites, Mica Creek, British Columbia. Am. Mineral. 64, 874-85.

GHENT, E. D. & STOUT, M. Z. 1981. Geobarometry and geothermometry of plagioclase-biotite-garnet-muscovite assemblages. Contrib. Mineral. Petrol. 76, 92-97.

GILETTI, B. J., MOORBATH, S. & LAMBERT, R. St. J. 1961. A geochronological study of the metamorphic complexes of the Scottish Highlands. Q. J. geol. Soc. London. 117, 233-72.

GOLDMAN, D. S. & ALBEE, A. L. 1977. Correlation of Mg/Fe partitioning between garnet and biotite with $^{18}O/^{16}O$ partitioning between quartz and magnetite. Am. J. Sci. 277, 750-67.

GOULD, D. 1965. Geochemical and mineralogical studies of the granite gneiss and associated rocks of western Ardgour, Argyll. Unpubl. Ph.D. Thesis, Univ. Edinburgh.

GROCOTT, J. 1979. Shape fabrics and superimposed simple shear strain in a Precambrian shear belt, W. Greenland. J. geol. Soc. London. 136, 471-87.

CUPPY, E. M. & SABINE, P. A. 1956. Chemical analyses of igneous rocks, metamorphic rocks and minerals. 1931-1954. Mem. geol. Surv. G. B.

HALLIDAY, A. N., AFTALION, M., VAN BREEMEN, O. & JOCELYN, J. 1979. Petrogenetic significance of Rb-Sr and U-Pb isotopic systems in the 400Ma old British Isles granitoids and their hosts. In HARRIS, A. L. et al. (1979), pp. 653-61.

HANSON, G. N. 1978. The application of trace elements to the petrogenesis of igneous rocks of granitic composition. Earth planet. Sci. Lett. 38, 26-43.

HARKER, R..I. 1954. Further data on the petrology of the pelitic hornfelses of the Carn Chuinneag-Inchbae region, Ross-shire, with special reference to the status of almandine. Geol. Mag. 91, 445-62.

HARRIS, A. L. & PITCHER, W. S. 1975. The Dalradian Supergroup. In HARRIS, A. L. et al. (1975), pp. 52-75.

HARRIS, A. L., SHACKLETON, R. M., WATSON, J. V., DOWNIE, C., HARLAND, W. B. & MOORBATH, S. 1975. A correlation of Precambrian rocks in the British Isles. Spec. Rep. geol. Soc. London. 6.

HARRIS, A. L., JOHNSON, M. R. W. & POWELL, D. 1978a. The orthotectonic Caledonides (Moines and Dalradians) of Scotland. Pap. geol. Surv. Can. 78-13, 79-85.

HARRIS, A. L., BALDWIN, C. T., BRADBURY, H. J., JOHNSON, H. D. & SMITH, R. A. 1978b. Ensialic basin sedimentation: the Dalradian Supergroup. In BOWES, D. R. & LEAKE (1978), pp. 115-38.

HARRIS, A. L., HOLLAND, C. H. & LEAKE, B. E. (eds.) 1979. The Caledonides of the British Isles - reviewed. Spec. Publ. geol. Soc. London. 8.

HARRISON, V. E. & MOORHOUSE, S. J. 1976. A possible early Scourian supracrustal assemblage within the Moine. J. geol. Soc. London. 132, 461-6.

HARRY, W. T. 1952. The migmatites and feldspar-porphyroblast rocks of Glen Dessary, Inverness-shire. Q. J. geol. Soc. London. 107, 137-158.

HARRY, W. T. 1953. The composite granitic gneiss of Western Ardgour, Argyll. Q. J. geol Soc. London. 109, 285-308.

HARVEY, C. O. & WILSON, A. D. 1956. In Summ. Prog. geol. Surv. G. B. for 1955, 58-63.

HELGESON, H. C., DELANEY, J. M., NESBITT, H. W. & BIRD, D. K. 1978. Summary and critique of the thermodynamic properties of rock-forming minerals. Am. J. Sci. 278A.

HIGGINS, A. C. 1967. The age of the Durine Member of the Durness Limestone formation at Durness. Scott. J. Geol. 3, 382-8.

McCLAY, K. R. 1981. What is a Thrust? What is a Nappe? Discussion with edited contributions from: J. G. DENNIS, R. A. PRICE, J. K. SALES, R. HATCHER, A. W. BALLY, W. J. PERRY, H. P. LAUBSCHER, R. E. WILLIAMS, D. ELLIOTT, D. K. NORRIS, D. W. HUTTON & T. EMMET. In McCLAY, K. R. & PRICE (1981), pp. 7-9.

McCLAY, K. R. & COWARD, M. P. 1981. The Moine Thrust Zone. An overview. In McCLAY, K. R. & PRICE (1981), pp. 241-60.

McCLAY, K. R. & PRICE, N. J. 1981. Thrust and nappe tectonics. Spec. Publ. geol. Soc. London. 9.

McINTYRE, D. B. 1952. The tectonics of the Beinn Dronaig area, Attadale. Trans. geol. Soc. Edinburgh. 15, 258-64.

MacQUEEN, J. A. & POWELL, D. 1977. Relationships between deformation and garnet growth in Moine (Precambrian) rocks of western Scotland. Bull. geol. Soc. Am. 88, 235-40.

MASON, B. 1966. Principles of geochemistry. John Wiley, New York.

MASUDA, A., NAKAMURA, N. & TANAKA, T. 1973. Fine structures of mutually normalised rare-earth patterns of chondrites. Geochim. cosmochim. Acta. 37, 239-48.

MEHNERT, K. R. 1968. Migmatites and the origin of granitic rocks. Elsevier, Amsterdam.

MERCY, E. L. P. 1963. The geochemistry of some Caledonian granitic and metasedimentary rocks. In JOHNSON, M. R. W. & STEWART (1963), pp. 189-215.

MENDUM, J. R. 1979. Caledonian thrusting in NW Scotland. In HARRIS, A. L. et al. (1979), pp. 291-7.

MICHEL-LEVY, A. 1893. Contribution à l'etude du granite de Flamanville et des granites français en général. Bull Serv. Carte geol. Fr. 5, (36):317.

MILLER, J. A. & BROWN, P. E. 1965. K-Ar age studies in Scotland. Geol. Mag. 102, 106-34.

MOORBATH, S. 1969. Evidence for the age of deposition of the Torridonian sediments of north-west Scotland. Scott. J. Geol. 5, 154-70.

MOORBATH, S. & TAYLOR, P. N. 1974. Lewisian age for the Scardroy mass. Nature. London. 250, 41-3.

MUNRO, M. 1965. Some structural features of the Caledonian granite complex at Strontian, Argyllshire. Scott. J. Geol. 1, 152-175.

MUNRO, M. 1973. Structures in the south-eastern portion of the Strontian granite complex, Argyllshire. Scott. J. Geol. 9, 99-108.

NEWTON, R. C., CHARLU, T. V. & KLEPPA, O. J. 1977. Thermochemistry of high-pressure garnets and clinopyroxenes in the system $CaO-MgO-Al_2O_3-SiO_2$. Geochim. cosmochim. Acta. 41, 369-77.

O'BRIEN, B. H. 1981. The structure and metamorphism of the Moine succession in Ardnamurchan, West Highlands, Scotland, with particular reference to the tectonics of the Morar Anticline. Unpubl. Ph.D. Thesis. Univ. Liverpool.

OLSEN, S. N. 1977. Origin of the Baltimore Gneiss migmatites at Piney Creek, Maryland. Bull geol. Soc. Am. 88, 1089-1101.

ORVILLE, P. M. 1972. Plagioclase cation exchange equilibria with aqueous chloride solution: results at 700°C and 2000bars in the presence of quartz. Am. J. Sci. 272, 234-72.

PANKHURST, R. J. 1979. Isotope and trace element evidence for the origin and evolution of Caledonian granites in the Scottish Highlands. In ATHERTON, M. P. & TARNEY (1979), pp. 18-33.

PARSON, L. M. 1979. The state of strain adjacent to the Great Glen Fault. In HARRIS, A. L. et al. (1979), pp. 287-9.

PEACH, B. N., HORNE, J, WOODWARD, H. B., CLOUGH, C. T., HARKER, A. & WEBB, C. B. 1910. The geology of Glenelg, Lochalsh and south-east part of Skye. Mem. geol. Surv. U. K.

PEACH, B. N., GUNN, W., CLOUGH, C. T., HINXMAN, L. W., CRAMPTON, C. B., ANDERSON, E. M. & FLETT, J. S. 1912. The geology of Ben Wyvis, Carn Chuinneag, Inchbae and the surrounding area. Mem. geol. Surv. U. K.

PHEMISTER, J. 1960. Scotland: The Northern Highlands. British Regional Geology. H. M. S. O.

PHILLIPS, F. C. 1971. The use of stereographic projection in structural geology. Edward Arnold, London.

PIASECKI, M. A. J. 1980. New light on the Moine rocks of the Central Highlands of Scotland. J. geol. Soc. London. 137, 41-59.

PIASECKI, M. A. J. & VAN BREEMEN, O. 1979. The 'Central Highland Granulites': cover-basement tectonics in the Moine. In HARRIS, A. L. et al. (1979), pp. 139-44.

PIDGEON, R. T. & JOHNSON, M. R. W. 1973. Isotopic evidence for 'early' events in the Moines. In PIDGEON, R. T., MacINTYRE, R. M., SHEPPARD, S. M. F. & VAN BREEMEN, O. (eds.). Geochronology and isotope geology of Scotland. Scottish Universities Research and Reactor Centre, East Kilbride.

PIDGEON, R. T. & JOHNSON, M. R. W. 1974. A comparison of zircon U-Pb and whole-rock Rb-Sr systems in three phases of the Carn Chuinneag granite, northern Scotland. Earth planet. Sci. Lett. 24, 105-12.

PIDGEON, R. T. & AFTALION, M. 1978. Cogenetic and inherited zircon U-Pb systems in granites: Palaeozoic granites of Scotland and England. In BOWES, D. R. & LEAKE (1978), pp. 183-220.

POOLE, A. B. 1966. The stratigraphy and structure of north-eastern Morar, Inverness-shire. Scott. J. Geol. 2, 38-53.

POOLE, A. B. & SPRING, J. S. 1974. Major structures in Morar and Knoydart, NW Scotland. J. geol. Soc. London. 130, 43-53.

POWELL, D. 1964. The stratigraphical succession of the Moine Schists around Lochailort (Inverness-shire) and its regional significance. Proc. Geol. Assoc. London. 75, 223-50.

POWELL, D. 1966. The structure of the south-eastern part of the Morar Antiform, Inverness-shire. Proc. Geol. Assoc. London. 77, 79-100.

POWELL, D. 1974. Stratigraphy and structure of the Western Moine and the problem of Moine orogenesis. J. geol. Soc. London. 130, 575-93.

POWELL, D., BAIRD, A. W., CHARNLEY, N. R. & JORDAN, P. J. 1981. The metamorphic environment of the Sgurr Beag Slide; a major crustal displacement zone in Proterozoic, Moine rocks of Scotland. J. geol. Soc. London, 138, 661-73.

POWELL, R. 1978. Equilibrium thermodynamics in petrology: an introduction. Harper & Row, London.

PURDY, J. W. & JAEGER, E. 1976. K-Ar ages on rock-forming minerals from the Central Alps. Mem. Inst. geol. mineral. Univ. Padova. 30.

QUINQUIS, H., AUDREN, C., BRUN, J. P. & COBBOLD, P. R. 1978. Intense progressive shear in Ile de Groix blueschists and compatibility with subduction or obduction. Nature, London, 273, 43-5.

RAMSAY, J. G. 1958a. Superimposed folding at Loch Monar, Inverness-shire and Ross-shire. Q. J. geol. Soc. London, 113, 271-307.

RAMSAY, J. G. 1958b. Moine-Lewisian relations at Glenelg, Inverness-shire. Q. J. geol. Soc. London, 113, 487-523.

RAMSAY, J. G. 1960. The deformation of early linear structures in areas of repeated folding. J. Geol. Chicago, 68, 75-93.

RAMSAY, J. G. 1962. The geometry and mechanics of formation of 'similar' type folds. J. Geol. Chicago, 70, 309-327.

RAMSAY, J. G. 1963. Structure and metamorphism of the Moine and Lewisian rocks of the North-West Caledonides. In JOHNSON, M. R. W. & STEWART (1963), pp. 143-75.

RAMSAY, J. G. 1967. Folding and fracturing of rocks. McGraw-Hill, New York.

RAMSAY, J. G. 1981. Tectonics of the Helvetic Nappes. In McCLAY, K. R. & PRICE (1981), pp. 293-309.

RAMSAY, J. G. & SPRING, J. 1962. Moine stratigraphy in the Western Highlands of Scotland. Proc. Geol. Assoc. London, 73, 295-322.

RATHBONE, P. A. 1980. Basement/cover relationships in the Moine Series of Scotland with particular reference to the Sgurr Beag Slide. Unpubl. Ph.D. Thesis, Univ. Liverpool.

RATHBONE, P. A. & HARRIS, A. L. 1979. Basement-cover relationships at Lewisian inliers in the Moine rocks. In HARRIS, A. L. et al. (1979), pp. 101-7.

READ, H. H. 1931. The geology of central Sutherland. Mem. geol. Surv. U. K.

READ, H. H. 1961. Aspects of Caledonian magmatism in Britain. Liverpool Manchester geol. J. 2, 653-683.

READ, H. H., PHEMISTER, J. & ROSS, G. 1926. The geology of Strath Oykell and lower Loch Shin. Mem. geol. Surv. U. K.

RICHARDSON, S. W. 1968. The petrology of the metamorphosed syenite in Glen Dessary, Inverness-shire. Q. J. geol. Soc. London, 124, 9-51.

RICHEY, J. E. & KENNEDY, W. Q. 1939. The Moine and sub-Moine series of Morar, Inverness-shire. Bull. geol. Surv. G. B. 2, 26-45.

ROBERTS, A. M. & HARRIS, A. L. In press. The Loch Quoich line - a limit of crustal reworking in the Northern Highlands of Scotland. J. geol. Soc. London.

ROBERTS, J. L. & TREAGUS, J. E. 1979. Stratigraphical and structural correlation between the Dalradian rocks of the SW and Central Highlands of Scotland. In HARRIS, A. L. et al. (1979), pp. 199-204.

SANDERSON, D. J. 1973. The development of fold axes oblique to the regional trend. Tectonophysics, 16, 55-70.

SHACKLETON, R. M. 1958. Downward-facing structures of the Highland Border. Q. J. geol. Soc. London, 113, 361-92.

SHEPHERD, J. 1973. The structure and structural dating of the Carn Chuinneag intrusion, Ross-shire. Scott. J. Geol. 9, 63-88.

SIEGEL, S. 1956. Nonparametric statistics for the behavioral sciences. McGraw-Hill, New York.

SMITH, D. I. 1979. Caledonian minor intrusions of the N Highlands of Scotland. In HARRIS, A. L. et al. (1979), pp. 683-97.

SOPER, N. J. 1971. The earliest Caledonian structures of the Moine thrust belt. Scott. J. Geol. 7, 241-7.

SOPER, N. J. & BROWN, P. E. 1971. Relationship between metamorphism and migmatisation in the northern part of the Moine nappe. Scott. J. Geol. 7, 305-25.

SOPER, N. J. & WILKINSON, P. 1975. The Moine Thrust and Moine Nappe at Loch Eriboll, Sutherland. Scott. J. Geol. 11, 339-59.

SOPER, N. J. & BARBER, A. J. 1982. A model for the deep structure of the Moine thrust zone. J. geol. Soc. London. 139, 127-38.

STEIGER, R. H. & JAEGER, E. 1977. Subcommission on geochronology: convention on the use of decay constants in geo- and cosmochronology. Earth planet. Sci. Lett. 36, 3 9-62.

STEVESON, B. G. 1968. A statistical analysis of chemical relationships in the Moine rocks of Lochailort. Unpubl. Ph.D. Thesis, Univ. Liverpool.

STEVESON, B. G. 1971. Chemical variability in some Moine rocks of Lochailort, Inverness-shire. Scott. J. Geol. 7, 51-60.

STOKER, M. S. 1980. The geology of eastern Ardgour, Argyllshire. Unpubl. Ph.D. Thesis, Univ. Liverpool.

STRACHAN, R. A. 1982. Tectonic sliding within the Moinian Loch Eil Division near Kinlocheil, W Inverness-shire. Scott. J. Geol. 18, 187 -203.

SUTTON, J. & WATSON, J. V. 1953. The supposed Lewisian inlier of Scardroy, central Ross-shire, and its relations with the surrounding Moine rocks. Q. J. geol. Soc. London. 108, 99-126.

SUTTON, J. & WATSON, J. V. 1954. On the status of certain Lewisian inliers. Trans. geol. Soc. Glasgow. 21, 480-502.

SUTTON, J. & WATSON, J. V. 1955. The structure and stratigraphical succession of the Moines of Fannich Forest and Strath Bran, Ross-shire. Q. J. geol. Soc. London. 110, 21-53.

SUTTON, J. & WATSON, J. V. 1962. An interpretation of Moine-Lewisian relations in Central Ross-shire. Geol. Mag. 99, 527-41.

TANNER, P. W. G. 1970. The Sgurr Beag Slide, a major tectonic break within the Moinian of the Western Highlands of Scotland. Q. J. geol. Soc. London. 126, 435-63.

TANNER, P. W. G. 1976. Progressive regional metamorphism of thin calcareous bands from the Moinian rocks of NW Scotland. J. Petrol. 17, 100-34.

TANNER, P. W. G., JOHNSTONE, G. S., SMITH, D. I. & HARRIS, A. L. 1970. Moinian stratigraphy and the problem of the central Ross-shire inliers. Bull. geol. Soc. Am. 81, 299-306.

TARNEY, J., SKINNER, A. C. & SHERATON, J. W. 1972. A geochemical comparison of major Archaean gneiss units from Northwest Scotland and East Greenland. Rep. 24th Int. geol. Congr. Montreal. 1, 162-74.

TARNEY, J. & SAUNDERS, A. D. 1979. Trace element constraints on the origin of cordilleran batholiths. In ATHERTON, M. P. & TARNEY (1979), pp. 90-105.

TARNEY, J., WEAVER, B. & DRURY, S. A. 1979. Geochemistry of Archaean trondhjemitic and tonalitic gneisses from Scotland and East Greenland. In BARKER, F. (ed.) Trondhjemites, dacites and related rocks, pp. 275-99. Elsevier, Amsterdam.

THOMPSON, A. B. 1974. Calculation of muscovite-paragonite-alkali feldspar phase relations. Contrib. Mineral. Petrol. 44, 173-94.

THOMPSON, A. B. 1976. Mineral reactions in pelitic rocks: II. Calculation of some P-T-X(Fe-Mg) phase relations. Am. J. Sci. 276, 425-54.

THOMPSON, A. B. & ALGOR, J. R. 1977. Model systems for anatexis of pelitic rocks. I. Theory of melting relations in the system $KAlO_2$ - $NaAlO_2$ - Al_2O_3 - SiO_2 - H_2O. Contrib. Mineral. Petrol. 63, 247-69.

THOMPSON, J. B. 1967. Thermodynamic properties of simple solutions. In ABELSON, P. H. (ed.). Researches in Geochemistry, 2, 340-61. John Wiley, New York.

TOBISCH, O. T. 1963. The structure and metamorphism of Moinian rocks in the Glen Cannich - Fasnakyle Forest area, Inverness. Unpubl. Ph.D. Thesis, Univ. London.

TOBISCH, O. T., FLEUTY, M. J., MERH, S. S., MUKHOPADHYAY, D. & RAMSAY, J. G. 1970. Deformational and metamorphic history of Moinian and Lewisian rocks between Strathconon and Glen Affric. Scott. J. Geol. 6, 243-65

TRACY, R. J. 1978. High grade metamorphic reactions and partial melting in pelitic schist, west-central Massachusetts. Am. J. Sci. 278, 150-78.

TRACY, R. J., ROBINSON, P. & THOMPSON, A. B. 1976. Garnet composition and zoning in the determination of temperature and pressure of metamorphism, central Massachusetts. Am. Mineral. 61, 762-75.

TREAGUS, J. E. & KING, G. 1978. A complete Lower Dalradian succession in the Schiehallion district, Central Perthshire. Scott. J. Geol. 14, 157-66.

TYLER, I. M. & ASHWORTH, J. R. 1982. Sillimanite-potash feldspar assemblages in graphitic pelites, Strontian area, Scotland. Contrib. Mineral. Petrol. 81, 18-29.

VAN BREEMEN, O., PIDGEON, R. T. & JOHNSON, M. R. W. 1974. Precambrian and Palaeozoic pegmatites of the Moines of northern Scotland. J. geol. Soc. London. 130, 493-507.

VAN BREEMEN, O., HALLIDAY, A. N., JOHNSON, M. R. W. & BOWES, D. R. 1978. Crustal additions in late Precambrian times. In BOWES, D. R. & LEAKE (1978), pp. 81-106.

VAN BREEMEN, O., AFTALION, M. & JOHNSON, M. R. W. 1979a. Age of the Loch Borrolan complex, Assynt, and late movements along the Moine Thrust Zone. J. geol. Soc. London. 136, 489-95.

VAN BREEMEN, O., AFTALION, M., PANKHURST, R. J. & RICHARDSON, S. W. 1979b. Age of the Glen Dessary syenite, Inverness-shire: diachronous Palaeozoic metamorphism across the Great Glen. Scott. J. Geol. 15, 49-62.

VAN DER MOLEN, I. & PATERSON, M. S. 1979. Experimental deformation of partially-melted granite. Contrib. Mineral. Petrol. 70, 299-318.

VIDALE, R. J. 1974. Vein assemblages and metamorphism in Dutchess County, New York. Bull. geol. Soc. Am. 85, 303-6.

WATSON, J. V. 1948. Late sillimanite in the migmatites of Kildonan, Sutherland. Geol. Mag. 85, 149-62.

WATSON, J. V. 1964. Conditions in the metamorphic Caledonides during the period of late-orogenic cooling. Geol. Mag. 101, 457-65.

WATSON, J. V. & DUNNING, F. W. 1979. Basement-cover relations in the British Caledonides. In HARRIS, A. L. et al. (1979), pp. 67-91.

WELLS, P. R. A. 1979. P-T conditions in the Moines of the Central Highlands, Scotland. J. geol. Soc. London. 136, 663-71.

WELLS, P. R. A. 1981. Reply to Winchester (1981). J. geol. Soc. London. 138, 206-9.

WELLS, P. R. A. & RICHARDSON, S. W. 1979. Thermal evolution of metamorphic rocks in the Central Highlands of Scotland. In HARRIS, A. L. et al. (1979), pp. 339-44.

WHITE, S. 1977. Geological significance of recovery and recrystallisation processes in quartz. Tectonophysics. 39, 143-70.

WILSON, D. 1975. Structure and metamorphism of the Ben Wyvis district, Ross-shire. Unpubl. Ph.D. Thesis, Univ. Edinburgh.

WILSON, D. & SHEPHERD, J. 1979. The Carn Chuinneag granite and its aureole. In HARRIS, A. L. et al. (1979), pp. 669-75.

WINCHESTER, J. A. 1974. The zonal pattern of regional metamorphism in the Scottish Caledonides. J. geol. Soc. London. 130, 509-24.

WINCHESTER, J. A. 1976. Different Moinian amphibolite suites in northern Ross-shire. Scott. J. Geol. 12, 187-204.

WINCHESTER, J. A. 1981. Discussion of Wells (1979). J. geol. Soc. London. 138, 205-6.

WINCHESTER, J. A., LAMBERT, R. St. J. & HOLLAND, J. G. 1981. Geochemistry of the western part of the Moine assemblage. Scott. J. Geol. 17, 281-94.

WINKLER, H. G. F. 1976. Petrogenesis of metamorphic rocks (4th ed.). Springer-Verlag, New York.

WONES, D. R. 1972. Stability of biotite: a reply. Am. Mineral. 57, 316-7.

WOOD, B. J. & FRASER, D. G. 1977. Elementary thermodynamics for geologists. Oxford University Press.

WYLLIE, P. J. 1977. Crustal anatexis: an experimental review. Tectonophysics. 43, 41-71.

YARDLEY, B. W. D. 1978. Genesis of the Skagit Gneiss migmatites, Washington, and the distinction between mechanisms of migmatisation. Bull. geol. Soc. Am. 89, 941-51.

YARDLEY, B. W. D., LEAKE, B. E. & FARROW, C. M. 1980. The metamorphism of Fe-rich pelites from Connemara, Ireland. J. Petrol. 21, 365-99.