# The Geology of the Moine rocks of the Loch Eil area, West Inverness-shire

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# CHAPTER 5: IGNEOUS ACTIVITY WITHIN THE LOCH EIL AREA

## 5.1. INTRODUCTION

The meta-igneous and igneous bodies of the Loch Eil area may be conveniently subdivided as follows:

| | | |
|---|---|---|
| (a) | West Highland granitic gneiss | Pre- to syn-tectonic |
| (b) | Amphibolite suites | |
| (c) | Felsic porphyrite-microdiorite-appinite suite | Late- to post-tectonic |
| (d) | Granite-vein complexes | |
| (e) | Agglomerate in vents | Permo-Carboniferous and/or Tertiary |
| (f) | Assorted camptonite, monchiquite and olivine basalt dykes | |

Only the pre- to syn-tectonic bodies are described in any detail in this chapter, since the petrology, field relations and timing of intrusion of the late- to post-tectonic and Permo-Carboniferous-Tertiary bodies are considered to be peripheral to the main objectives of this study. Details of the felsic porphyrite-microdiorite-appinite suite are given by Dearnley (1967) and Smith (1979). The granite-vein complexes of the Loch Eil area are described in detail by Fettes and MacDonald (1978); the approximate western margin of their "Banavie Complex", as mapped during the present study, is indicated in Enclosure 2. A small monchiquite vent on Druim Fada is described by Hartley and Leedal (1951).

## 5.2. WEST HIGHLAND GRANITIC GNEISS

### 5.2.1. General remarks

The granitic gneiss bodies which crop out between Strontian and Glen Moriston are collectively known as the West Highland granitic gneiss (Johnstone 1975). Two members of the suite are present within the Loch Eil area: the southern body is commonly known as the 'Ardgour granitic gneiss', and the northern is termed, in this study, the 'Gulvain granitic gneiss'. Published work (Johnstone et al. 1969) regards the formation of the members of the West Highland Granitic Gneiss as broadly synchronous, and the two granitic gneiss bodies of the Loch Eil area, which are similar in their petrological and structural characteristics, are therefore referred to in this study as 'the granitic gneiss'. Both granitic gneiss bodies are separated from the Loch Eil Division by the Druim Na Saille Pelite. The eastern margin of the Ardgour Granitic Gneiss was mapped from the Cona River as far N as Glen Dubh Lighe, and the southwestern margin of the Gulvain Granitic Gneiss was mapped from Na-h-Uamachan northeastwards to the western slopes of Gulvain. The full extent of the outcrop of these granitic gneiss bodies is depicted in Fig. 4.

### 5.2.2. Field description

The granitic gneiss forms massive, rounded, pale-coloured outcrops and in hand speciman is seen to consist of quartz, plagioclase and potash feldspar, and biotite. It is easily distinguishable from the adjacent Druim Na Saille Pelite which has a greater proportion of biotite to quartz and feldspar. Texturally the granitic gneiss is a medium to coarse-grained gneiss. A penetrative gneissosity is typically defined by the alternation of thin (1-5 millimetres), elongate (1-3 centimetres) quartzo-feldspathic aggregates with thin (1-2 millimetres) discontinuous folia of biotite (Plate 44 ). Occasionally, this gneissosity may be considerably coarser (e.g. in the Allt a Choire Reidh, NM 98558530) where the quartzo-feldspathic aggregates may be up to 1 centimetre thick and laterally continuous for distances of up to 10 centimetres.

The granitic gneiss commonly contains numerous quartzo-feldspathic pegmatites, which impart a migmatitic appearance. The pegmatites are broadly concordant on outcrop scale with the dominant gneissosity within the granitic gneiss (Plate 45), but in detail occasionally cross-cut this fabric. They are distinguishable from the enclosing granitic gneiss by their greater proportion of quartz and feldspar relative to biotite. They are typically lenticular in form, and range up to 20 centimetres in length and 2 centimetres in thickness. Discrete biotite-rich selvedges, 2-3 millimetres thick, are occasionally present at the margins of the pegmatites; at some localities these selvedges are thicker and display gradational contacts with the granitic gneiss.

### 5.2.3. Petrology

There is little to add to the excellent petrographic descriptions of the granitic gneiss provided by Dalziel (1963) and Harry (1953). Microscopic study shows that the granitic gneiss is composed principally of quartz, microcline, oligoclase ($An_{14}$-$An_{28}$) and biotite, with accessory apatite, zircon, iron ores and sphene.

Average grain-size within the quartzo-feldspathic aggregates is 1-2 millimetres. Quartz is present either as individual xenoblastic grains or as lenticular aggregates: grain boundaries are often intricately sutured. Undulose extinction is locally common. Microcline and oligoclase are intergrown with the quartz, both occurring as xenoblastic, approximately equidimensional grains. Microcline grains often display well-developed gridiron twinning; oligoclase is distinguished by albite twinning. Myrmekitic intergrowths between the two are locally common. Discrete bands composed of idioblastic laths of biotite (1-3 millimetres long) enclose the quartzo-feldspathic aggregates. Of the accessory minerals, zircon and the iron cores (mainly pyrite and magnetite) are concentrated within the biotite-rich bands; apatite and sphene do not show any systematic distribution.

### 5.2.4. Relationships with the Druim Na Saille Pelite

Contacts between the granitic gneiss and the Druim Na Saille Pelite are well exposed in the Allt a Choire Reidh (NM 98558515) and the Allt a Choire Chruinn (NM 92707530), and patchily exposed on the western slopes of Glas Bheinn. The main features of these contacts may be summarised thus:

(a) The transition from granitic gneiss to semi-pelitic gneiss occurs over a distance of 1-2 metres.

(b) There is no discordance of foliation across the contacts.

(c) No visible increase in the level of strain is apparent within either the granitic gneiss or the Druim Na Saille Pelite as their contacts are approached.

Plate 44: Characteristic fabric of the granitic gneiss. (Ardgour granitic gneiss, NM 934799).

Plate 45: Concordant quartzo-feldspathic pegmatites within the granitic gneiss. The pegmatite to the left of the compass is isoclinally folded. (Ardgour granitic gneiss, NM 934799).


PENTAX

(d) The semi-pelitic gneisses adjacent to the contact do not display any more or less migmatisation than elsewhere.

(e) The granitic gneiss was not observed to enclose any bands or lenticles of semi-pelitic gneiss or amphibolite.

### 5.2.5. Structure of the granitic gneiss

The gneissosity within the granitic gneiss is entirely concordant with the main foliation within the adjacent pelitic and psammitic gneisses of the Glenfinnan Division. This fabric is very occasionally folded by small-scale isoclines (Plate 45) which are difficult, however, to assign to a specific deformational event. The apparent lack of minor folds within the granitic gneiss is partly a function of the small proportion mapped of the total outcrop of the granitic gneiss. It also seems probable that a large proportion of the regional strain was accommodated during successive deformational events by the lithologically heterogeneous Druim Na Saille Pelite and Gulvain Psammitic Gneiss, resulting in a relative lack of minor structures within the granitic gneiss.

### 5.2.6. Origin of the granitic gneiss

There would appear to be three broad alternatives with regard to the origin of the granitic gneiss:

(a) Firstly, that the granitic gneiss is the result of potash metasomatism and the in situ replacement of Moine rocks, as envisaged by Harry (1953) and Dalziel (1966).

(b) Secondly, that the granitic gneiss represents slices of allochthonous basement which were tectonically emplaced into the Moine rocks, as suggested by Harris (in discussion of Winchester 1974).

(c) Finally, that the granitic gneiss represents a series of pre- to syn-tectonic granitic melts intruded into the rocks of the Glenfinnan Division (Mercy 1963; Gould 1966; Aftalion and Van Breemen 1980).

Each of these alternatives will be discussed in turn.

5.2.6.1. The granitic gneiss as a metasomatic replacement of Moine rocks

Harry (1953) envisaged that the granitic gneiss and the enclosing semi-pelitic gneisses formed as part of an evolutionary sequence of regional metasomatism. This sequence began with an early phase of Na, Ca and Mg metasomatism, forming the semi-pelitic oligoclase-rich gneisses, and was followed by "a main wave of granitisation" during which Si and K metasomatism at the lowest structural levels led to the formation of the granitic gneiss. Dalziel (1966) broadly supported Harry's conclusions, and considered that the granitic gneiss was possibly a result of the recrystallisation of metasediments under sillimanite-almandine sub-facies conditions during the uprise of potassic-rich volatiles from zones of partial melting at lower levels. He suggested that the steepening of the axes of second phase folds along the Loch Quoich Line led to increased upward tectonic transport, aiding the uprise of potassic-rich volatiles. In support of their hypothesis that the granitic gneiss formed as a result of the passive metasomatic replacement of host paragneisses, both Harry and Dalziel pointed to:

(a) The broad parallelism of foliation between the granitic gneiss and the enclosing metasediments: the gneissosity within the granitic gneiss was interpreted as a relic of the schistosity formed in the metasediments preserved as a gneissose foliation after mimetic coarsening of the fabric, regional migmatisation and subsequent metasomatism.

(b) The many concordant inclusions of pelitic gneiss and hornblende schist within the granitic gneiss: these were interpreted as zones of the host paragneisses which had evaded replacement.

(c) The transitional contacts between the granitic gneiss and the enclosing metasediments, this transition occurring over distances ranging from less than 1 metre to several metres. Both workers considered that this supported a metasomatic origin for the granitic gneiss, since they assumed that an intrusive magmatic body would display sharp contacts.

(d) The broadly symmetrical disposition of stratigraphic units about a major antiform containing in its core the granitic gneiss, which appears to be concordant with the stratigraphy and to have formed at the same stratigraphic level. Dalziel suggested that the granitic gneiss occupies the same stratigraphic level as the Ben Tuim Striped Group, which crops out further W, and appears to have replaced this unit.

(e) Dalziel (1963) also concluded that the zircon grains in the granitic gneiss were virtually identical in their anhedral morphology to those of the enclosing metasediments, and this was interpreted as a further indication that the granitic gneiss was a metasomatic replacement. Following the work of Poldervaart (1950,

1956) who suggested that euhedral zircons were characteristic of magmatic granites, Dalziel proposed that the anhedral shape of the zircons in the granitic gneiss was a direct indication that the granitic gneiss was not magmatic in origin.

Much of the above evidence put forward by Harry and Dalziel in support of their hypothesis is, however, equivocal, and cannot be considered to be conclusive of a purely metasomatic origin for the granitic gneiss. The parallelism of foliation between the granitic gneiss and the metasediments could equally well be the result of deformation superimposed upon an intrusive body concordant with its enclosing country rocks. Similarly, metasedimentary inclusions within the granitic gneiss might be interpreted as xenolithic blocks incorporated within an intrusive melt and subsequently rotated into parallelism with the gneissosity during deformation. Purely amphibolitic inclusions could possibly represent primary residua of a granite melt rather than country rock (Pidgeon and Aftalion, 1978). The view that an intrusive magmatic body would display sharp contacts with its country rock is overly simplistic. The transitional contacts linking the granitic gneiss with the adjacent semi-pelitic gneisses could be analogous to those bordering many intrusive bodies, such as the Strath Halladale granite. In this case, a transitional zone of 'injection migmatites' links the granite with its Moine country rock ( Lintern et al, in press) and the granite appears to have been intruded into already warm metasediments with the result that the granite does not display a chilled margin and the Moine rocks were not hornfelsed. Furthermore, the validity of Dalziel's conclusions concerning the relationship between the shape of zircons and their origin is questioned by the work of Pankhurst and Pidgeon (1976), who show that zircons from the intrusive Ben Vuirich granite are anhedral and bear a close morphological resemblance to those of the Ardgour granitic gneiss.

5.2.6.2. The granitic gneiss as allochthonous basement

Harris (in discussion of Winchester, 1974) suggests that the granitic gneiss might represent slices of granitic basement which had been tectonically emplaced into Moine cover, presumably along zones of high strain. It has already been indicated (p.106 ) that there is no visible increase in the level of strain within either the granitic gneiss or the Druim Na Saille Pelite as their contacts are approached: this suggests that these contacts are not tectonic. However, it is important to emphasise that this observation only relates to the upper boundary of the granitic gneiss; if the suggestion that the granitic gneiss occupies the core of a major antiform is correct (Dalziel, 1966, p.142), then the lower, and unexposed, boundary could still be tectonic.

5.2.6.3. The granitic gneiss as pre- to syn-tectonic intrusions

It has already been indicated that the field relationships of the granitic gneiss are equivocal, and a reinterpretation of previous work involving the granitic gneiss as pre- to syn-tectonic intrusions could be equally valid. This viewpoint has been adopted by several workers. Mercy (1963, p.214), in a geochemical study, commented on the "quite remarkable magmatic features" of the granitic gneiss and suggested that it might represent a "distinct kind of magma emplaced amongst high grade metamorphic rocks". Gould (1966) subsequently demonstrated that the granitic gneiss had a uniform minimum melt composition, and considered that it originated as a granitic melt which was formed at a depth not far below the present exposure level. Aftalion and Van Breeman (1980) broadly agreed with Gould's interpretation and argued, on the basis of isotopic evidence, that the granitic melt was largely derived from the extensive melting and assimilation of Moine metasediments at a lower tectonic level. They also draw attention to the presence of parallel growth twins in the zircons within the granitic gneiss, which might also indicate that the granitic gneiss had been through a magmatic stage (Jocelyn and Pidgeon 1974).

The following additional observations might also suggest that this interpretation is broadly correct:

(a) The granitic gneiss is both texturally and petrologically homogeneous over large areas. It seems unlikely that this degree of homogeneity could result from the metasomatism of such a heterogeneous assemblage of metasediments as the Ben An Tuim Striped Group, which comprises a series of coarsely-interbanded semi-pelitic gneisses, psammites and quartzites. A more reasonable explanation of this feature is that the granitic gneiss is the deformed equivalent of an intrusive granite sheet.

(b) Although Dalziel maintained that the Ardgour granitic gneiss was concordant with the local Moine stratigraphy, this does not appear to be entirely correct. The western margin of the

Ardgour granitic gneiss is in contact with both the Druim Na Saille Pelite and the Ben An Tuim Striped Group (Fig. 4), both of which thin markedly north-northeastwards. This cannot be wholy explained by the distribution of major folds in the area (Dalziel, 1966, p.142). There is no evidence that the western boundary of the granitic gneiss is tectonic, since it displays the same features as the eastern boundary (pers. obsv.). This discordance between the granitic gneiss and the stratigraphy seems most likely,therefore, to represent an original intrusive contact.

### 5.2.7. Age of the foliation within the granitic gneiss

This is considered in relation to the deformational sequence established in Chapter 4. Two observations are of particular importance:

(a) The gneissose fabric within the granitic gneiss is concordant with the main foliation within the Druim Na Saille Pelite and Gulvain Psammitic Gneiss throughout the area, regardless of structural setting. It is considered likely that the two foliations are of the same age.

(b) The granitic gneiss is migmatitic. Since there is only one major phase of migmatisation apparent within the adjacent metasediments (p. 137), which was synchronous with the production of their main foliation, it seems likely that they were both migmatised at the same time. This provides a second indication that both the granitic gneiss and the metasediments of the Glenfinnan Division have undergone a broadly similar tectono-metamorphic history.

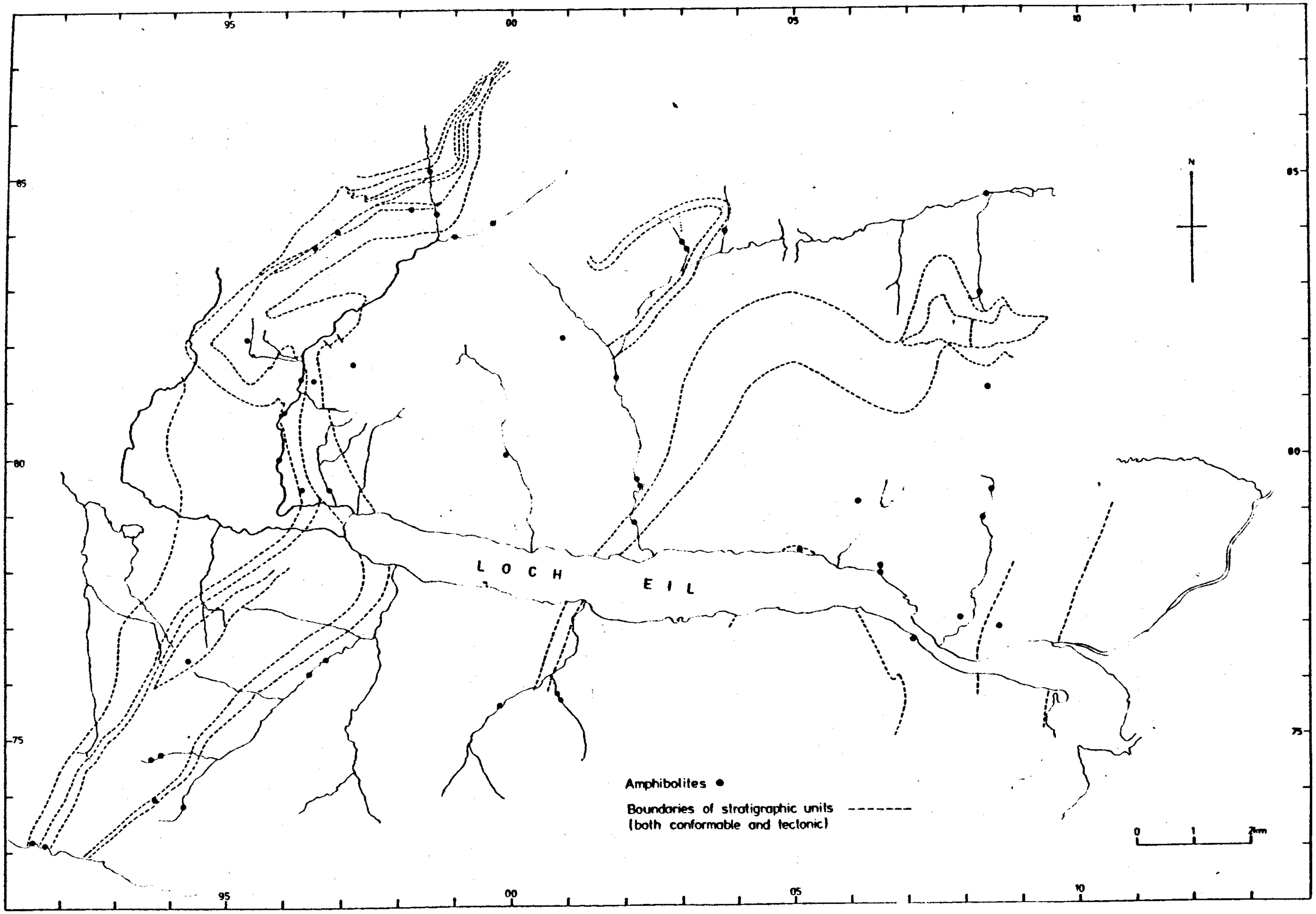
### 5.2.8. Conclusions

The most plausible explanation for the origin of the granitic gneiss is that it originated as granitic melt intruded into the rocks of the Glenfinnan Division. Evidence adduced in support of the interpretation that it represents the result of the in situ metasomatic replacement of host paragneisses is equivocal, and a magmatic origin is more consistent with both the observed field evidence and the geochemical and isotopic evidence produced by previous workers. Intrusion was either pre- or syn-tectonic with respect to the generation of the main foliation within the adjacent metasediments of the Glenfinnan Division. The transitional contacts which link the granitic gneiss and the Druim Na Saille Pelite, together with the lack of any relict hornfels zone adjacent to the granitic gneiss, are both indications that granitic melt may have been emplaced syn-tectonically into already warm rock.

## 5.3. AMPHIBOLITES

### 5.3.1. General remarks

Small pods and sheets of amphibolite have been recorded at 45 localities, randomly distributed throughout the area (Fig.33 ). Their appearance in the field, petrology, relationships with the surrounding metasediments and structural history will each be described in turn.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### 5.3.2. Field description

The majority of amphibolites have a sheet-like form, but are rarely traceable laterally for any great distance. They range in size from relatively small sheets or pods, 10-30 centimetres thick (Plate 46), to larger bodies (Plate 47) which may attain thicknesses of 3-4 metres. Exceptionally, amphibolite bands may be as thin as 2-3 millimetres (e.g. NN 071768). Certain amphibolites appear to have served as the locus for the intrusion of quartz veins which are locally common within some bodies (e.g. NN 071768). These quartz veins range in thickness from 2-10 millimetres, and may be laterally continuous for distances of up to 1 metre; other quartz veins may take the form of small detached elliptical globules only a few millimetres across.

In hand specimen the amphibolites are dark green to black, and consist of a medium to coarse-grained assemblage of hornblende, plagioclase and quartz, with occasional flakes of biotite. A few amphibolites display a compositional banding which results from the alternation of mafic and felsic components on a scale of 2-3 millimetres; generally, however, this feature is rare. All possess a penetrative foliation arising from the preferred alignment of hornblende prisms. Certain larger bodies ($>$1 metre thick) appear to have been comparitively resistant to deformation, and display highly foliated medium-grained margins which contrast markedly with their weakly foliated and coarse-grained centres.

### 5.3.3. Petrology

The dominant mineral assemblage is hornblende + plagioclase + quartz $\pm$ biotite $\pm$ garnet, with sphene a common accessory. Most

Plate 46: Small amphibolite pod partially enclosed by migmatitic semi-pelitic gneiss. (Druim Na Saille Pelite, NM 781915).

Plate 47: Amphibolite sheet within semi-pelitic schist. (Cona Glen Psammite, NN 079770).

specimens contain 70-80% hornblende and 20-30% felsic minerals.

Hornblende is typically present as aligned idioblastic or sub-idioblastic laths which are 0.5-2.0 millimetres long. Most are pale to dark-green and display good cleavage. They are usually aggregated in a series of anastomosing streams which enclose irregular patches of quartz and feldspar (Plate 48). Evidence for a second generation of hornblende is apparent in the occasional presence of xenoblastic hornblende which overgrows at high angles the earlier and dominant fabric. Those hornblendes occuring in the less deformed centres of larger amphibolites are typically poorly-oriented, sub-idioblastic or xenoblastic laths, 1-2 millimetres long (Plate 49). They frequently have ragged, ill-defined margins, and enclose numerous small quartz grains.

Plagioclase and quartz are intergrown in lenticular aggregates which are entirely surrounded by hornblende prisms. Both occur as xenoblastic, approximately equidimensional grains, most of which are 0.1-0.5 millimetres in diameter. Grain boundaries are usually slightly curved. Most quartz grains display straight extinction; undulose extinction is locally common, however. Plagioclase is distinguished by albite twinning and varies in composition from andesine $An_{45}$ to oligoclase $An_{27}$. Sericitisation is generally rare.

Biotite occurs both as an 'early' phase, which is intergrown with hornblende, (Plate 50) and as a 'late' phase which cross-cuts the hornblende fabric. Where it is present as an late phase, it is represented

Plate 49: Coarse hornblende fabric typical of the less deformed centres of larger amphibolites (NN 079770).

Plate 48: Hornblende laths aligned in S1 within amphibolite. (Nm 96307935).

Scale bar represents 0.1mm.

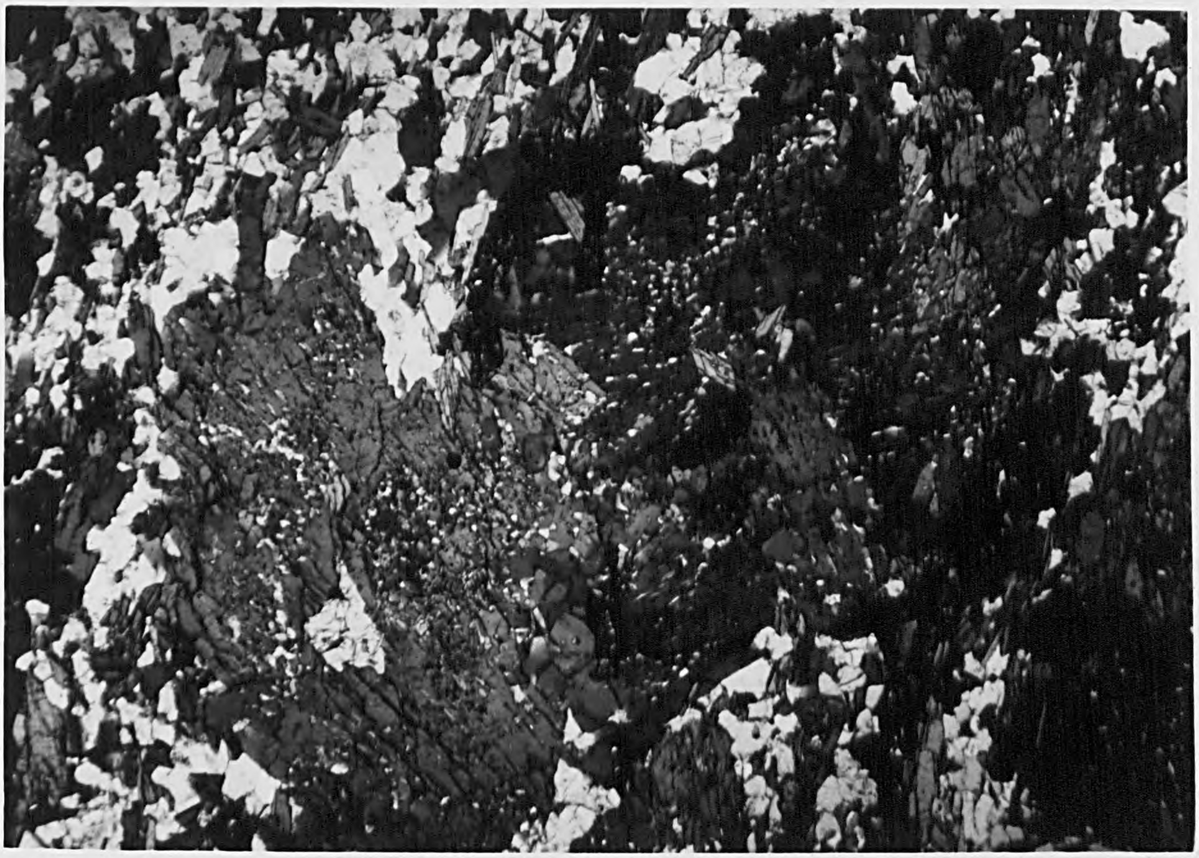


Plate 50: Biotite and hornblende intergrown in S1 within amphibolite (NM 96307935).

Plate 51: 'Late' phase xenoblastic biotite overgrowing S1 hornblende fabric within amphibolite (NM 93857470).

Scale bar represents 0.1 mm.

by large randomly oriented xenoblastic laths, 1-3 millimetres long (Plate 51).

Garnet has only been identified in one specimen, where it occured as a sub-idioblastic grain 3 millimetres across, which included numerous small grains of sphene and quartz.

Sphene is a common accessory, typically occuring as sub-idioblastic grains 0.1-0.3 millimetres long. In most specimens it is irregularly distributed; occasionally, however, it forms elongate aggregates 3-4 millimetres long.

Epidote and pyrite are both occasionally present, as randomly distributed xenoblastic grains or aggregates 0.1-0.5 millimetres across.

### 5.3.4. Relationships with the enclosing metasediments

The margins of most amphibolites are entirely concordant with the dominant foliation within the enclosing metasediments. Contacts with the metasediments are sharp and generally planar. Discordant contacts with the enclosing metasediments are displayed at three localities:

(a) Near Fassfern at (NN 02157885) the planar margin of an amphibolite cross-cuts the S1 foliation within the enclosing Druim Fada Quartzite at approximately $5^{o}$.

(b) South of Loch Eil at (NN 071768) thin amphibolite bands (1-2 centimetres thick) diverge slightly from the S1 foliation within the Cona Glen Psammite over a distance of 8 metres. In detail these bands may have highly irregular margins, with numerous small veins of amphibolite penetrating the psammite (Fig. 34).

(c) Clear evidence of two generations of amphibolite is forthcoming in the Allt a Choire Reidh (NM 98558510), where 'late' amphibolite sheets are markedly discordant both with respect to the foliation within the adjacent Gulvain Psammitic Gneiss, and to a series of 'early' amphibolites (Fig.35 ).

### 5.3.5. Origin of the amphibolites

The amphibolites may have originated either as thin bands of sediment which were particularly rich in iron and magnesia, or as pre- to syn-tectonic intrusives. The possibility that they may also represent thin lavas or tuffs must also be considered (Winchester 1976). In the absence of preserved igneous textures within the amphibolites, only those bodies which are discordant to the foliation within the adjacent metasediments can be positively identified as representing original igneous intrusives (section 5.3.4. above). The following observations suggest, however, that the remainder, and indeed majority, of amphibolites may also have originated as thin concordant sheets of basic magma:

(a) The contacts between the amphibolites and adjacent metasediments are always sharp, and not gradational as might be expected if the amphibolites represent original sedimentary units.

(b) The presence of sphene as a common accessory indicates that the parent rocks were characterised by a high proportion of $TiO_2$. According to Leake (1964) this suggests that the amphibolites are more

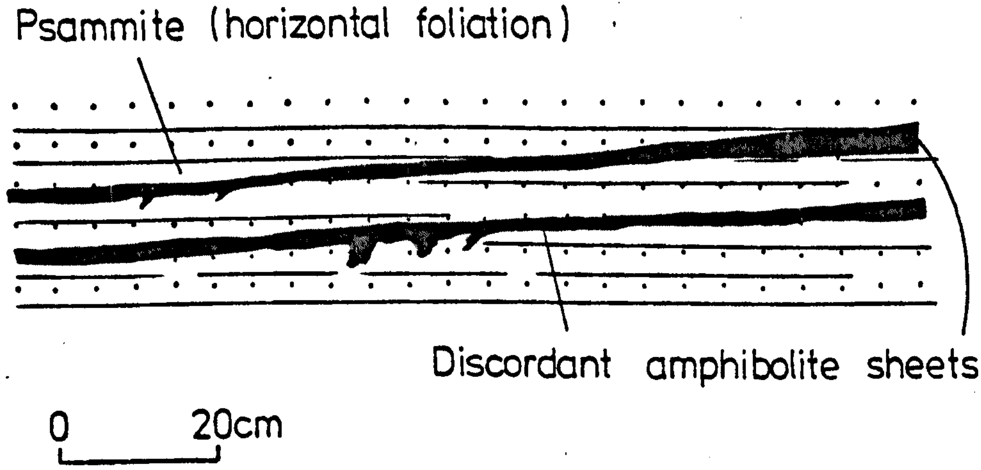



Figure 34: Amphibolite - country-rock relationships at (NN 071768).

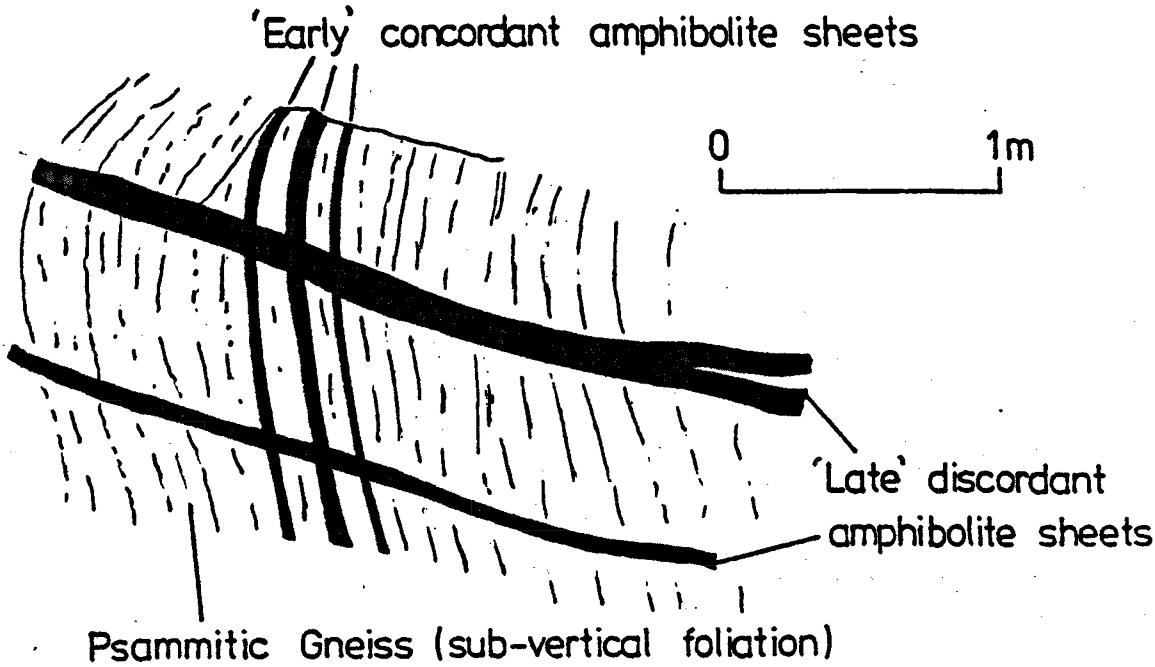



Figure 35: 'Early' and 'late' amphibolite sheets within the Gulvain Psammitic Gneiss (NM 98558510).

likely to be meta-igneous in origin than metasedimentary.

### 5.3.6. Age of intrusion of the amphibolites

This is considered with respect to the deformational sequence described in Chapter 4. The following observations are important:

(a) The penetrative foliation within those amphibolites recorded from the Loch Eil Division is in all cases parallel to S1 within the enclosing metasediments. At two localities within the Loch Eil Division (NN 071768 and NN 0557685) amphibolites are deformed by D1 folds, and at each the foliation within the amphibolites is axial planar to these structures. These observations clearly indicate that the amphibolites within the Loch Eil Division were emplaced either prior to, or synchronous with, D1.

(b) The penetrative foliation within the concordant amphibolites recorded from the Glenfinnan Division is similarly parallel to the main foliation within the enclosing metasediments. Since these amphibolites are identical in form and petrology to those within the Loch Eil Division, it is assumed that they were all emplaced at the same time as one suite. However, it is uncertain whether these amphibolites acquired their foliation as a result of the early structural events within the Glenfinnan Division (p. 97 ), or whether their emplacement entirely postdated these events and they were foliated during D1.

(c) The 'late' amphibolite sheets in the Allt a Choire Reidh (section 5.2.4.(c) above) only develop a weak hornblende fabric which is parallel to the axial planes of local D2 folds. This suggests that their intrusion postdated D1, but predated D2.

(d) Following their emplacement and initial deformation, amphibolites were deformed during D2 and presumably also D3-D5. Amphibolites are deformed by D2 minor folds in the Cona River (NM 91457315) (Plate 52) and in the Allt Beithe (NM 93857470). At both localities the foliation within the amphibolites is folded by these structures, and there is little growth of hornblende parallel to the axial planes of the D2 folds.

## 5.3.7. Conclusions

The present study demonstrates the existence of at least two amphibolite suites within the Moine rocks of the Loch Eil area:

(a) The first, and most extensive, suite was injected into the metasediments of the Glenfinnan and Loch Eil Divisions as a series of thin concordant sheets of basic magma. Emplacement of these into the Loch Eil Division was either pre- or syn-tectonic with respect to D1.

(b) Members of the second suite are restricted in their occurrence to a locality in the Allt a Choire Reidh where they cross-cut members of the first suite. Intrusion of these bodies postdated D1, but predated D2.

These findings are broadly in accord with the work of Peacock (1977) and Smith (1979), both of whom recognise two intrusive suites

Plate 52: Thin amphibolite band and enclosing migmatitic semi-pelitic gneiss deformed by a tight D2 fold. (Druim Na Saille Pelite, NM 915781).

of amphibolites within Moine rocks of W. Inverness-shire. The first suite recognized by them is represented by a series of thin concordant sheets which appear to have been involved in all the deformational events which affect their host rocks. According to Smith, the members of this suite are largely confined to the Loch Eil Division, within which they are extensively distributed. This suite is considered to be directly analagous to the first suite described above from the Loch Eil area. Both workers also recognise a second suite of amphibolites which cross-cut the earliest fabrics within their host rocks. Smith (1979, Fig. 5) indicates that the occurrence of members of this suite is restricted to the 'Loch Quoich Line', and it is considered likely that this suite corresponds to the second suite described above from the Loch Eil area.

# CHAPTER 6: METAMORPHIC HISTORY OF THE LOCH EIL AREA

## 6.1. INTRODUCTION

The metamorphic history of the Loch Eil area has been studied with reference to various techniques. Indications of both the grade of metamorphism, and areal variations in grade, are provided by the examination of the mineral assemblages within the metasediments and the amphibolites, and also by textural variations within the metasediments. The relative timing of mineral growth may be related to successive deformational phases.

## 6.2. METAMORPHIC HISTORY OF THE LOCH EIL DIVISION

### 6.2.1. Evidence of metamorphic grade

#### 6.2.1.1. Pelitic and quartzo-feldspathic assemblage

This assemblage is characteristic of the majority of the metasediments and comprises biotite + muscovite + garnet + quartz + andesine-oligoclase. Garnet is present throughout the area and its presence suggests that the area has been subject to metamorphism of at least low amphibolite facies (Turner 1981, p.209). The precise determination of metamorphic grade using the index minerals of the Barrovian zonal sequence is difficult since occurrences of such minerals as staurolite, kyanite and sillimanite are rare within the Moine Succession, even where alternative evidence suggests that the pressure and temperature conditions would have favoured the growth of such minerals in rocks of a suitable composition. The comparative scarcity of these minerals appears to be related to the chemical composition of Moine rocks (Kennedy 1948; Winchester 1974). Nevertheless, fibrolitic sillimanite has been recorded from 3 localities in the W of the area (p.43 ) and its presence suggests that, at least locally, metamorphism attained upper amphibolite facies (Turner op cit, p. 209).

### 6.2.1.2. Basic assemblage

This assemblage is characteristic of the amphibolites and comprises hornblende + plagioclase ($An_{25}$) + quartz ± biotite. The presence of the key mineral pair hornblende-plagioclase ($An_{>25}$) is regarded as diagnostic of metamorphism of amphibolite facies (Turner op cit, p. 366). Anorthite content displays a systematic variation within the area, progressively rising from high oligoclase ($An_{27-30}$) in the E, to high andesine ($An_{40-45}$) in the W. Numerous workers have demonstrated a general increase in anorthite content with rising metamorphic grade (e.g. Wenk & Keller 1969), and it seems likely that the observed variation corresponds to the transition from low to mid-amphibolite facies. Biotite is present as an 'early' phase intergrown with hornblende in both the E and W of the area, and its occurrence does not, therefore, appear to be controlled by metamorphic grade (cf. Wiseman 1934), but more probably relates to local variations in rock composition (Turner op cit, p. 366).

### 6.2.1.3. Calcareous assemblages

These are characteristic of the calc-silicates, and comprise:

1 (a) Hornblende + biotite + andesine + garnet + zoisite + quartz.

(b) Hornblende + andesine + garnet + quartz ± biotite ± zoisite ± calcite.

(c) Hornblende + bytownite + garnet + quartz ± zoisite ± biotite ± clinozoisite.

(d) Pyroxene + bytownite + garnet + quartz ± hornblende ± zoisite ± clinozoisite.

2 Hornblende + epidote + andesine-oligoclase + quartz ± garnet ± zoisite.

The Arnipol-type calc-silicates of assemblage 2 are less responsive to changes in metamorphic grade than those calc-silicates of assemblage 1 (Winchester 1972; Powell *et al.* 1981), and will not be considered in this chapter. Numerous workers (e.g. Winchester 1974; Powell *et al.* *op cit*) have demonstrated that the mineral assemblages of calc-silicates are dependant not only on metamorphic grade, but also on the whole rock $CaO/Al_2O_3$ ratio. Thus Powell *et al.* (*op cit*), indicates that calc-silicates with $CaO/Al_2O_3$ ratios of 0.353 and above contain amphibole and/or pyroxene in environments where pelitic rocks contain staurolite, kyanite and sillimanite. Where alumino-silicates are absent on a regional scale, biotite and zoisite characterise calc-silicates with $CaO/Al_2O_3$ ratios of less than 1.003, and may be considered to be indicative of garnet grade conditions. A further useful indicator of metamorphic grade is plagioclase composition, since the anorthite content of plagioclase tends to show a progressive increase with rising grade (Winchester 1974; Tanner 1976; Powell *et al.* *op cit*). All the calc-silicates of assemblage 1 above have $CaO/Al_2O_3$ ratios of between 0.353 and 1.003 (Winchester pers. comm.) and they are therefore of considerable use in assessing metamorphic grade.

The ubiquitous presence of primary hornblende or pyroxene within assemblage 1 suggests metamorphism of amphibolite facies. A more detailed study of the mineral assemblages indicates that the sequence 1(a) to (d) is a prograde one developed entirely within the confines of the amphibolite facies. The lowest grade assemblages are 1(a) and (b) where the co-existence of andesine and hornblende indicates metamorphism of low amphibolite facies (Tanner _op cit_). The additional presence of primary biotite and zoisite in 1(a) suggests that this is of slightly lower metamorphic grade than 1(b). The presence of the mineral pair hornblende-bytownite (assemblage 1(c)) is generally considered to suggest metamorphism of mid-amphibolite facies (Tanner _op cit_ p.116), and is equated with the appearance of kyanite in pelitic metasediments. The highest grade assemblage is 1(d) where the co-existence of pyroxene and bytownite indicates metamorphism of upper amphibolite facies (Winchester 1974; Tanner _op cit_), and has been correlated with the occurrence of sillimanite in pelitic metasediments.

Given the geographic distribution of the members of assemblage 1 (p.50 ), the calc-silicate mineralogy suggests that metamorphic grade is lowest in the eastern part of the area, progressively rising westwards.

### 6.2.1.4. Textural evidence within the metasediments

The following textural observations are important:

(a) The size of the quartz grains within semi-pelitic and pelitic lithologies is relatively uniform across the area (0.2-0.4 millimetres) but shows a marked increase in the area to the W of the North Garvan River and Fionn Lighe (0.3-0.8 millimetres), and on Na-h-Uamachan these lithologies are locally gneissose. This suggests a rise in metamorphic grade along the western margin of the area.

(b) Quartz grains within calc-silicates show a similar variation in size from 0.2-0.4 millimetres in the E of the area, to 0.5-1.0 millimetres in the W. Garnets within the calc-silicates progressively change in form and size from xenoblastic or sub-idioblastic grains, 0.2-0.7 millimetres in diameter, (assemblage 1(a)), to spongy ill-defined networks, 2-3 millimetres in diameter (assemblage 1(c)(d)). These textural changes are identical to those recorded by Tanner (op cit) from prograde sequences in the Kinloch Hourn area, and supports the mineralogical evidence which indicates that grade increases westwards.

(c) The nature of the microscopic fabrics within the quartzites also provides important information relating to the grade of metamorphism. Specimens obtained from the Druim Fada Quartzite, the Stronchreggan Mixed Assemblage, and thin quartzite units within the Glen Garvan Psammite on Beinn an-t-Sneachda are characterised by the following features:

(i) Average grain size is 0.1-0.3 millimetres.

(ii) Quartz-quartz and quartz-plagioclase boundaries are slightly curved, and triple point junctions which meet at high dihedral angles are common.

(iii) Quartz grains are mostly equant and polygonal.

(iv) Micas occur between quartz-quartz boundaries.

These features are broadly consistent with the attainment of textural equilibrium (Spry 1969, p. 159; Wilson 1973, p.60), and the formation of a stable primary recrystallised matrix structure. Wilson (op cit p.44) suggests that such a fabric is characteristic of metamorphism of upper greenschist to mid-amphibolite facies.

In contrast, specimens examined from the Kinlocheil Cross-Bedded and Banded Quartzite display:

(i) A considerable variation in quartz grain size from 0.1-2.0 millimetres. The size of plagioclase grains remains relatively stable at 0.1-0.4 millimetres.;

(ii) Quartz-quartz boundaries are highly intricate and meet at a variety of dihedral angles. Plagioclase grains are frequently embayed by lobate growths of quartz.

(iii) Quartz grains are irregular and complex in form.

(iv) Micas and other accessories are no longer confined to between quartz grains, but tend to be either partially or completely enclosed by large quartz grains.

The major features of these textural variations are illustrated in Plates 53 and 54.

Plate 53: 'Equilibrium-type' quartzite fabric:

(i) Average grain size of 0.1-0.3 millimetres.

(ii) Quartz grains are approximately equant and polygonal.

(iii) Grain boundaries are relatively simple in form.

(Druim Fada Quartzite, NN 04008225).

Plate 54: Recrystallised and coarsened quartzite fabric:

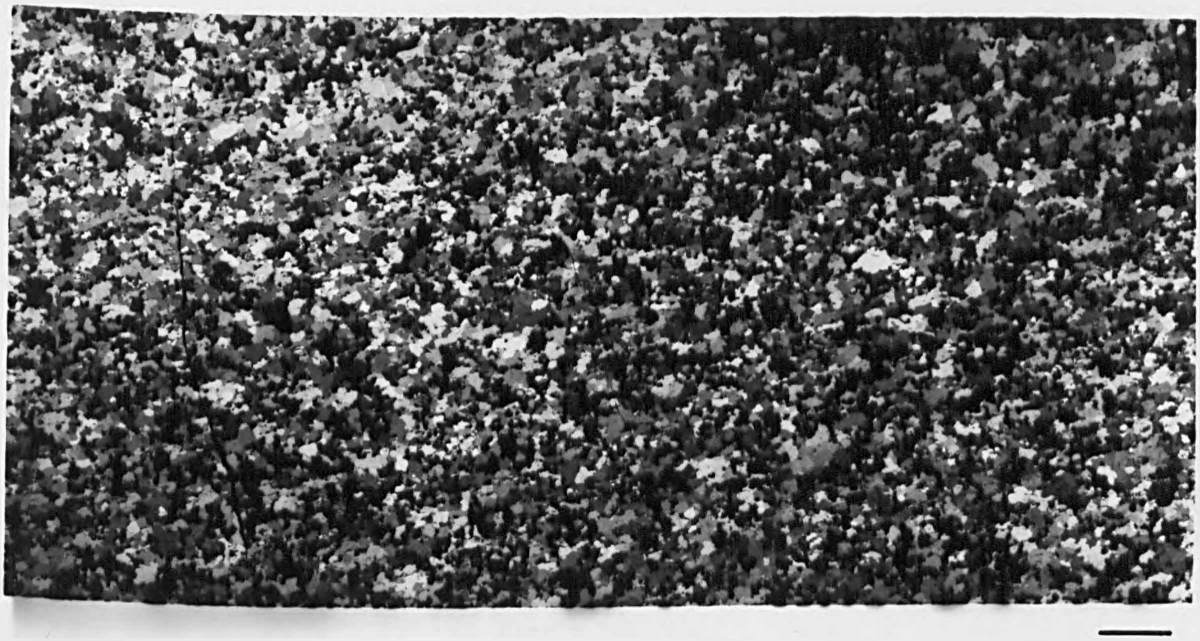
(i) Considerable variation in grain size from 0.1-2.0 millimetres.

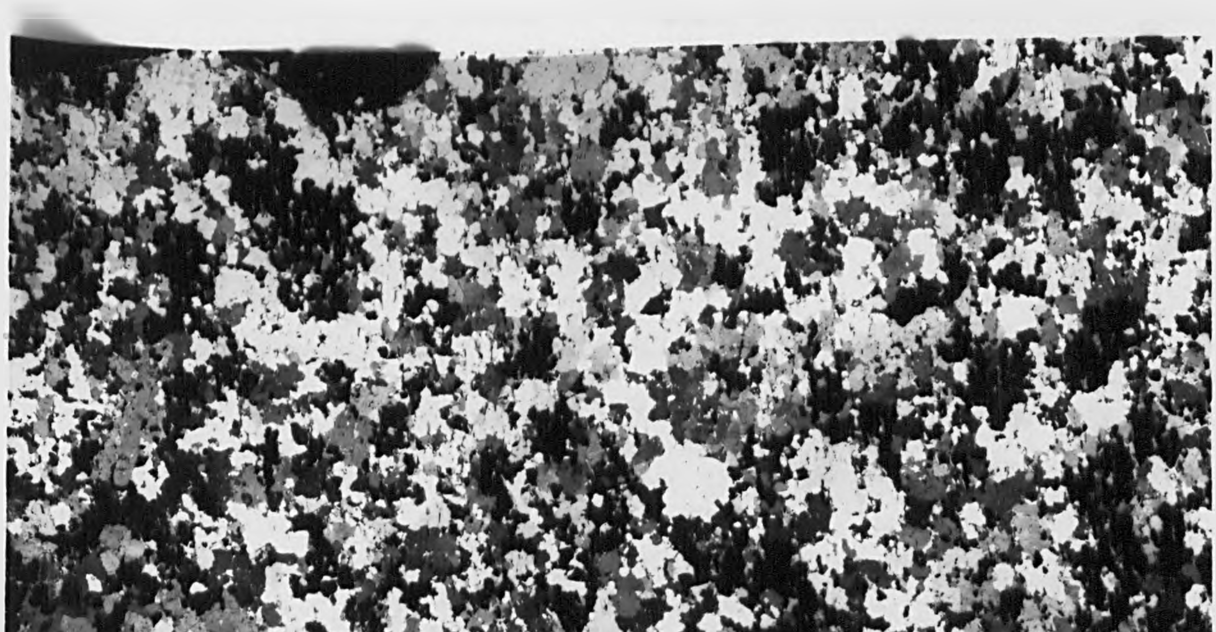
(ii) Quartz grains are irregular and complex.

(iii) Grain boundaries are typically intricate.

(Kinlocheil Banded Quartzite, NM 962835).

Scale bar represents 1mm.





Wilson (op cit p.61) suggests that these features result from the secondary recrystallization of a stable primary recrystallized matrix structure. This involves the abnormal coarsening of a few quartz grains which become very large and absorb both the finer polygonal primary recrystallised matrix and envelop micas. The form of the coarsened grains is typically irregular, and their boundaries are usually complex. According to Wilson (op cit p. 44) such a fabric is indicative of upper amphibolite facies metamorphism.

This variation in quartzite fabrics is therefore suggestive of an increase in metamorphic grade along the western margin of the Loch Eil Division. The textural variations observed within the quartzites are paralleled to a limited extent within psammitic lithologies. The Basal Psammite, and Glen Garvan Psammite on Sron an-t-Sluichd and Na-h-Uamachan, display some of the features attributed to secondary recrystallisation, although they are not as well developed as within the quartzites. It seems likely that the increased presence of feldspar and mica inhibited the secondary recrystallisation and growth of quartz (Wilson op cit p. 55).

6.2.1.5. Conclusions concerning metamorphic grade

All the available mineralogical and textural information suggests that the Loch Eil Division has been subject to metamorphism of amphibolite facies, and that grade increased from E to W.

A detailed subdivision of the Loch Eil Division on the basis of metamorphic grade is difficult, but it is possible, however, to broadly subdivide the area into two (Fig. 36). The eastern area,

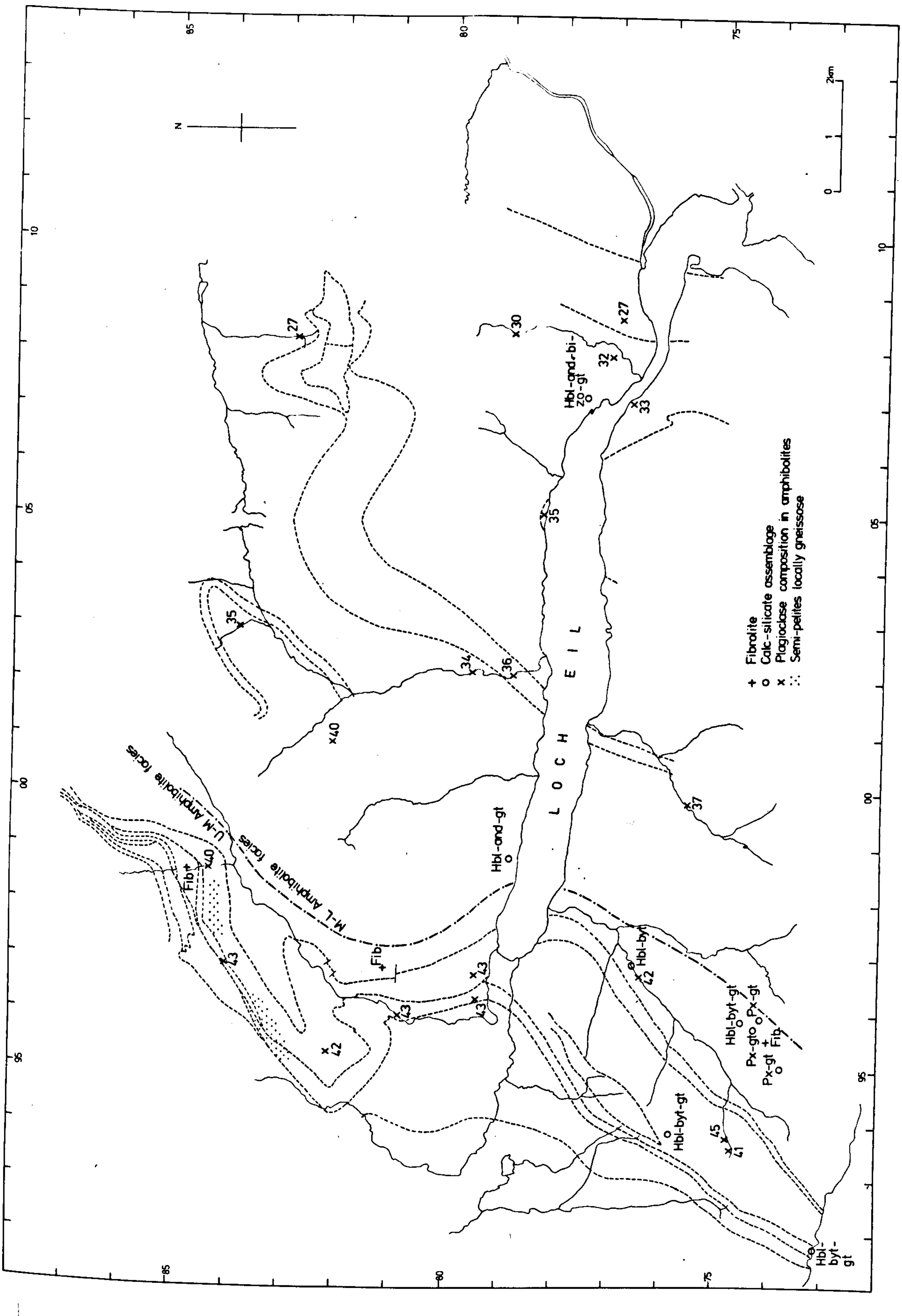



Figure 36: Variations in metamorphic grade within the Loch Eil Division.

## 6.2.2. Relationship of mineral growth to deformation

### 6.2.2.1. Pelitic and quartzo-feldspathic assemblage

Biotite and muscovite occur as oriented laths which define the regional S1 foliation. Crystal growth was therefore largely syn-tectonic with the first period of deformation. Subsequent development is reflected in the local growth of some biotite and muscovite laths parallel to the axial planes of D2 minor folds and D3 crenulations.

Both are also present as disoriented xenoblastic porphyroblasts, the dating of which is more difficult (late-tectonic, D3-D4?). The occurrence of muscovite porphyroblasts might also be attributed to local K-metasomatism. Both garnet and plagioclase porphyroblasts are enveloped by S1 and may locally be elongate in the plane of S1 (Plate 55 ). This suggests that their growth largely predated the first phase of deformation; there is no evidence to suggest that renewed growth occurred at any other time.

Alkali-feldspar-quartz aggregates described earlier (Ch.3, p. 43 , Plate 14) are similarly elongate in the plane of S1, but it is difficult to establish their relationship to the S1 mica fabric. They may therefore have formed either before or during the main phase of D1 deformation, during which they were flattened, or entirely postdate D1 and have grown mimetically along S1. The last alternative is the less likely since their host rocks are not notably micaceous and there is unlikely to have been any great directional control on their growth.

Plate 55: Elongate garnet enveloped by S1 mica fabric. (Cona Glen Psammite, NN 079770, plane polarised light).

Scale bar represents 0.1mm

Plate 56: Faserkiesel aligned in S1. (Glen Garvan Psammite, NM 954740).


PENTAX

The faserkiesel which contain fibrolitic sillimanite display a strong preferred orientation parallel to S1 (Plate 56). Where they are deformed by D2 minor folds, the majority of faserkiesel are crinkled (Plate 57) although some recrystallise parallel to the axial-planes of the D2 structures. This suggests two phases of growth: the earlier, and more important, of the two is difficult to time precisely but is likely to have been broadly syn-tectonic with respect to D1; the second, and more localised, phase is synchronous with D2.

The occurrence of chlorite may be almost entirely attributed to the late-stage largely post-tectonic (D5?) retrogression of biotite.

6.2.2.2. Basic assemblage

The alignment of hornblende and biotite defines the S1 foliation within the amphibolites. Crystal growth was therefore largely syn-tectonic with respect to D1. Subsequent growth of hornblende has only been detected at Kinlocheil (NM 96307935) where second phase hornblendes are aligned broadly parallel to the axial planes of D2 and D3 folds. In this case it is considered more likely that growth occurred during D2. Post-D1 growth of biotite porphyroblasts as a result of the retrogression of hornblende is difficult to date precisely (late-tectonic, D3-D4?).

6.2.2.3. Calcareous assemblage

The timing of pyroxene growth is difficult to establish since

Plate 57: D2 folding of a semi-pelitic band containing numerous faserkiesel, most of which are either crinkled or remain aligned in S1, although some recrystallise in S2. (Glen Garvan Psammite, Nm 954740).

Plate 58: Folding of a calc-silicate pod containing assemblage 1(a) by a tight D1 structure. The biotite-hornblende fabric within the calc-silicate is aligned in S1. (Cona Glen Psammite, NN 075774).


PENTAX

its relationship to S1 within the calc-silicates is not seen. However, it seems likely that its growth occurred during D1 at the same time as the growth of sillimanite within adjacent semi-pelites.

The calc-silicates of assemblages 1(a)(b)(c) carry a penetrative fabric resulting from the alignment of hornblende prisms. Within 1(a) this is enhanced by the intergrowth of biotite (Plate 21). This fabric is parallel to S1 within the enclosing metasediments, and is occasionally seen to be axial-planar to D1 folds (Plate 58). This suggests that the growth of primary hornblende and biotite in these assemblages is largely synchronous with D1. The second generation of hornblende present has a distinctive fibrous form and results in 1(c) from the recrystallisation of primary hornblende, and in 1(d) from the retrogression of pyroxene. It is suggested that this occurred during D2. A second generation of biotite present in 1(b)(c) results from the retrogression of primary hornblende, and is difficult to date precisely (late-tectonic D3-D4?).

Garnets are locally enveloped by hornblende lying in S1, and it seems likely that garnet growth largely preceded D1. There is no textural evidence to suggest that the growth of zoisite and clinozoisite may be attributed to 'late' retrogression of 'early' mineral assemblages, and it is considered that both grew prior to and during D1.

#### 6.2.2.4. Conclusions concerning timing of mineral growth

The observations relating to this are summarised in Table 10,

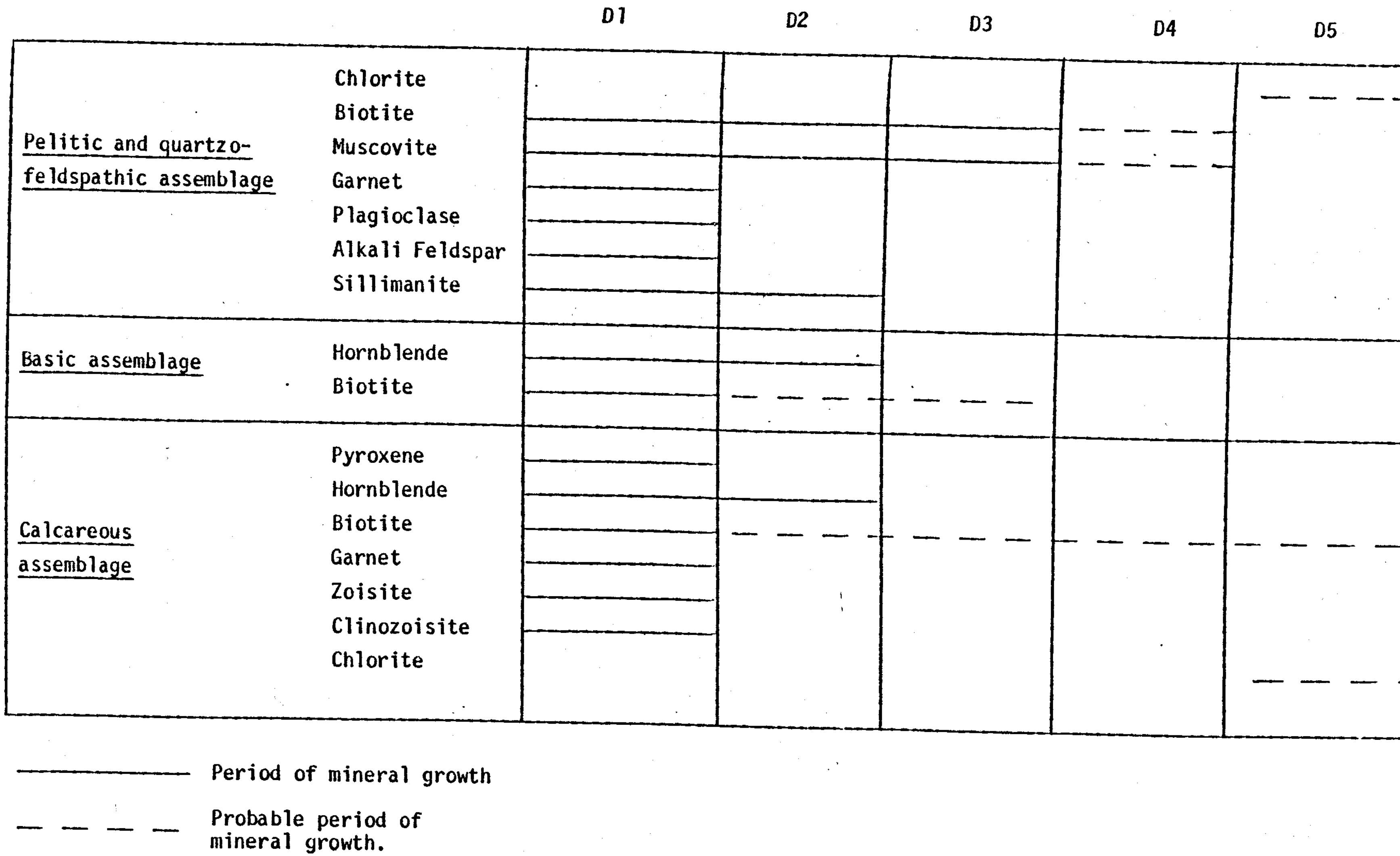



Table 10: Relationship of mineral growth to deformation

and clearly indicate that the Loch Eil Division has been subjected to two major episodes of metamorphism which were broadly synchronous with D1 and D2 respectively. Subsequent metamorphism is of little significance.

### 6.2.3. Summary

The metamorphic history of the Loch Eil Division is summarized in Table 11 and comprises:

(a) A major metamorphic episode (M1) which was synchronous with D1. The metamorphic fabric of the Loch Eil Division was largely acquired during this event, which was of upper to mid-amphibolite facies grade along the western margin of the area, diminishing in grade eastwards with decreasing tectonic level to mid- to low amphibolite facies.

(b) A second metamorphic episode (M2) which was synchronous with D2. Limited renewed growth of M1 assemblages occurred, together with the slight retrogression of pyroxene to hornblende in calc-silicates. Metamorphic grade was in all likelihood still high (mid- to upper amphibolite facies) but slightly lower than during M1.

(c) Localised late-tectonic (D3-D4) growth (M3) of biotite and muscovite in the pelitic and quartzo-feldspathic assemblage, and retrogression of hornblende to biotite in amphibolites and calc-silicates. M1 and M2 assemblages remained otherwise stable during M3, the grade of which is likely to have been at least upper greenschist facies.

| META-MORPHIC EPISODE | TIMING RELATIVE TO STRUCTURAL HISTORY | DIAGNOSTIC ASSEMBLAGES: WEST OF AREA: Pelitic and quartzo-feldspathic | WEST OF AREA: Basic | WEST OF AREA: Calcareous | EAST OF AREA: Pelitic and quartzo-feldspathic | EAST OF AREA: Basic | EAST OF AREA: Calcareous | PROBABLE METAMORPHIC GRADE |
|---|---|---|---|---|---|---|---|---|
| M1 | Syn-D1 | sillimanite + garnet + biotite + muscovite | hornblende + biotite + plag($An_{40-45}$) | bytownite + pyroxene (hornblende) + garnet | garnet + biotite + muscovite | hornblende + biotite + plag ($An_{27-40}$) | hornblende + andesine + garnet | Upper to Mid-Amphibolite Facies along the western margin of the area, decreasing eastwards to Mid to Low-Amphibolite Facies. |
| | | Also: 1. Local gneissose fabrics<br>2. Complex quartzite fabrics | | | | | | |
| | | FORMATION OF THE DOMINANT MINERAL ASSEMBLAGES AND METAMORPHIC FABRIC OF THE LOCH EIL DIVISION | | | | | | |
| M2 | Syn-D2 | Localised renewed growth of M1 sillimanite, hornblende, biotite and muscovite.<br>Retrogression of M1 pyroxene to hornblende. | | | | | | Amphibolite Facies (overall, probably slightly lower than during M1, although locally grade attained Upper Amphibolite Facies) |
| M3 | Late-tectonic (D3-D4?) | Localised renewed growth of biotite and muscovite in pelitic and quartzo-feldspathic assemblages. Retrogression of hornblende to biotite in basic and calcareous assemblages. The majority of mineral growth is unoriented. | | | | | | Upper Greenschist Facies (minimum) |
| M4 | Largely post-tectonic (D5?) | Retrogression of biotite to chlorite (unoriented) | | | | | | Lower Greenschist Facies |

Table 11: Summery of the metamorphic history of the Loch Eil Division

(d) Largely post-tectonic (D5?) retrogression (M4) of biotite to chlorite within the pelitic and quartzo-feldspathic and calcareous assemblages during metamorphism of low greenschist facies.

## 6.3. METAMORPHIC HISTORY OF THE GLENFINNAN DIVISION

This is investigated in rather less detail than that of the Loch Eil Division, mainly as a result of the smaller area mapped.

### 6.3.1. Evidence of metamorphic grade

The following considerations are important:

(a) The ubiquitous presence of garnet within the Druim Na Saille Pelite suggests metamorphism of at least low amphibolite facies. Dalziel (1963), however, reports the presence of aligned sillimanite within the Druim Na Saille Pelite in Cona Glen and Glen Dubh Lighe, and this indicates that grade is in reality somewhat higher than suggested above.

(b) The dominant assemblage within amphibolites is hornblende + plagioclase ($An_{40-45}$).

(c) Calc-silicates examined from (NM 954793) contain hornblende and bytownite, and have $CaO/Al_2O_3$ ratios of approximately 0.4 (Winchester, pers. comm.). Dalziel (op cit) also reports the presence of pyroxene in calc-silicates in Glen Dubh Lighe.

(d) The Druim Na Saille Pelite and Gulvain Psammitic Gneiss are coarse-grained, and possess well-developed gneissose fabrics. The latter is inferred to have resulted from the small-scale segregation of pelitic and quartzo-feldspathic components during metamorphism and deformation; much segregation may have been mimetic after bedding.

(e) Microscopic fabrics within thin quartzite bands, and to a lesser extent the psammitic gneisses, are identical to those described from the Kinlocheil Cross-Bedded and Banded Quartzites (p. 127).

These considerations suggest that the Druim Na Saille Pelite and Gulvain Psammitic Gneiss have been subject to metamorphism of mid- to upper amphibolite facies.

6.3.2. Relationship of mineral growth to deformation

(a) The alignment of biotite and muscovite laths within pelitic and psammitic lithologies, and hornblende prisms within calc-silicates, defines the main foliation within the Druim Na Saille Pelite and Gulvain Psammitic Gneiss (p. 97 ). Crystal growth was therefore syn-tectonic with respect to the formation of this fabric and, by inference, phase 1 folding (p. 97 ). Garnets are enveloped by the mica fabric and their growth therefore largely predated its formation. From the descriptions provided by Dalziel (1963) it seems likely that the growth of sillimanite in pelitic gneiss, and pyroxene in calc-silicates, also accompanied the formation of this fabric. There is no information relating to the grade of metamorphism during phase 2 folding.

(b) Subsequent development is restricted to:

(i) The localised growth of biotite and muscovite parallel to the axial planes of D1 (Plate 41) and D2 minor folds (correlated with M1 and M2 respectively).

(ii) The retrogression of hornblende to biotite in amphibolites and calc-silicates, and the growth of biotite porphyroblasts in pelitic lithologies (correlated with M3).

(iii) The retrogression of biotite to chlorite (correlated with M4).

### 6.3.3. Timing and mode of Formation of migmatites

The Druim Na Saille Pelite, in common with large parts of the Glenfinnan Division, is frequently migmatitic (p.47 and Plates 1, 17, 18 and 52) and it is often suggested that this is an indication that the Glenfinnan Division has undergone partial to complete anatexis (e.g. Winchester 1974, p.515).

The formation of migmatitic lits in rocks cannot easily be related to a particular phase of folding, since if the lits are folded, they may (i) predate the fold (ii) form syntectonically, or (iii) post-tectonically by selective reconstitution of layers of appropriate composition. However, migmatitic leucosomes are entirely concordant with the main foliation within the Druim Na Saille Pelite, and are deformed by D1 (Plate 42) and D2 (Plates 1 and 52) minor folds. There is no direct indication that migmatisation occurred during the formation of any of these folds, as might be suggested by the presence of axial planar lits, and it therefore seems most likely that migmatisation was broadly contemporaneous with the production of the main foliation within the Druim Na Saille Pelite.

Migmatites may be produced by a variety of mechanisms, either in an 'open' system by means of igneous injection or external metasomatism, or in a 'closed' system involving anatexis or metamorphic segregation. Whilst some petrologists favour an anatectic origin for the majority of migmatites (e.g. Mehnert 1968; Ashworth 1975, 1977; Winckler 1979), others point to the importance of hydrothermal processes in their genesis (White 1966; Yardley 1978).

It seems unlikely that the migmatitic features observed were formed as a result of the introduction of material by either igneous injection or external metasomatism in an 'open' system. There are none of the cross-cutting relationships which might be expected if any substantial amount of quartzo-feldspathic material had been externally derived, and igneous injection by itself does not seem to be a reasonable explanation for regional migmatisation of the Glenfinnan Division. The occurrence and relative proportions of the different minerals in the migmatitic rocks are closely comparable to those of their presumed equivalents present as somewhat less migmatised enclaves, and there is no obvious petrological indication that migmatisation was the result of the metasomatic introduction of any 'foreign' mineral phase.

It is more likely, therefore, that migmatisation occurred within a 'closed' system as a result of anatexis or metamorphic segregation. Yardley (1978) discusses the criteria by which it might be possible to distinguish between the products of these two mechanisms, and with regard to this the following observations derived from the migmatitic rocks of the Druim Na Saille Pelite would appear to be of importance:

(a) Migmatitic leucosomes are largely trondjhemitic in composition, and only rarely contain K-feldspar. This is inconsistent with an origin involving anatexis, since melting studies on water-saturated metasediments have shown that K-feldspar crystallises from the first-formed partial melts of a wide range of metamorphic rocks (von Platen 1965). The K-feldspar is derived from the breakdown of biotite, leaving such phases as cordierite or garnet in the restite.

(b) The albite content of plagioclase within leucosomes is comparable to that within the palaeosomes. This feature is also inconsistent with an anatectic model. Melting studies on plagioclase and plagioclase-bearing systems suggest that the albite component of plagioclase is strongly fractionated into the co-existing melt phase, so that after recrystallisation the plagioclase of the granitic leucosomes formed by anatexis should be 10-40% more sodic than that of the restite (Drake 1976). The uniformity of plagioclase composition is, however, more consistent with an origin involving metamorphic segregation which would not incorporate any fractionation of plagioclase.

(c) Leucosomes are often closely-spaced (Plate 59). Yardley (op cit p.942) argues that such morphology is more likely to be a result of metamorphic segregation, where only a small amount of fluid need be present at any one time, than anatexis where the total volume of leucosome is molten at one time. In the latter case the rock will be mechanically weakened and agmatisation is likely to occur, involving fracturing and rotation of country rock blocks in zones of extensive melting. There is, however, no trace of such phenomena in the study area.

(d) Well-developed melanosomes (e.g. Plate 18) are the exception rather than the rule, and basified margins to leucosomes are more typically either poorly-developed or entirely absent (Plate 17). This is more consistent with an origin involving metamorphic segregation than differentiation of a molten neosome during anatexis (Yardley op cit p.943).

Although the validity of some of Yardley's criteria has been challenged (e.g. Johannes and Gupta 1982), taken together these considerations suggest that metamorphic segregation is likely to have been the major agent of migmatisation, and there is no compelling evidence to indicate that anatexis was responsible. However, given that the grade of metamorphism during migmatisation is likely to have been, at least locally, within the upper amphibolite facies, and thus within the melting field for water-saturated sediments, it is conceivable that a degree of partial melting may have occurred contemporaneously with metamorphic segregation.

Plate 59: Closely-spaced leucosomes within a migmatitic semi-pelitic gneiss (Druim Na Saille Pelite, NM 95157910).


0
6 cm.

## 6.4. CORRELATION WITH PREVIOUS WORK

This study is in broad agreement with the work of Stoker (1980) concerning both the grade and timing of successive metamorphic events within the Loch Eil Division. There are, however, certain differences between the metamorphic history proposed in this study, and that suggested by Dalziel (1963), the most important being that whilst Dalziel proposes that peak metamorphism occurred during 'F2' ('D2' of this study), both the present author and Stoker (op cit) consider that peak regional metamorphism occurred during D1, and that grade progressively declined thereafter.

Winchester (1974) proposes that sillimanite-grade rocks occur in the E of the Loch Eil area adjacent to the Great Glen Fault. This suggestion is based on the presence of hornblende and bytownite in a calc-silicate collected from Banavie Quarry, subsequent reinvestigation during the present study has failed to confirm this, and only Arnipol-type calc-silicates have been obtained from this locality. Some of these are texturally complex (p. 57 and Plate 24), and contain prominent aggregates of hornblende and epidote. Winchester (pers. comm.) has suggested that these might be the result of the wholesale M2 retrogression of M1 pyroxene. However, there is no positive trace of any preserved pyroxene, and such extreme retrogression would certainly be unusual since retrogressive effects elsewhere in the area are generally negligible. Moreover, there does not seem to be any obvious reason why metamorphic grade during M1 might progressively decline from W to E with decreasing tectonic level, and then abruptly increase. Accordingly, it seems more likely to the present author that the hornblende-epidote aggregates referred to above are part of a prograde M1 Arnipol-type assemblage, and there is therefore little direct evidence that grade in this area is as high as has been suggested.

## 6.5. CONCLUSIONS

These may be summarised as follows:

(a) Four metamorphic events (M1-M4) have been recognised within the Loch Eil Division. M1 and M2 are the most important of these, and were broadly synchronous with D1 and D2 respectively.

(b) The dominant metamorphic fabric of the Loch Eil Division is a reflection of peak metamorphism during M1, which was of mid- to upper amphibolite facies in the W of the area, progressively declining to low to mid-amphibolite facies in the E.

(c) Thereafter, metamorphic grade gradually diminished during the sequence M2-M4, the effects of which are limited to localised renewed growth and retrogression of M1 assemblages.

(d) The metamorphic fabric of the Glenfinnan Division is largely a result of mid- to upper amphibolite facies metamorphism and widespread migmatisation, both of which accompanied the formation of the dominant foliation within the Druim Na Saille Pelite and Gulvain Psammitic Gneiss. Whether this metamorphic event is correlatable with M1 within the Loch Eil Division, or whether it is an entirely separate and much earlier event depends upon the temporal relationship between 'pre-D1 structural elements' and 'D1-D5' (p.96 ).

# CHAPTER 7: SEDIMENTOLOGY OF THE LOCH EIL DIVISION

## 7.1. INTRODUCTION

### 7.1.1. General remarks

There has been virtually no systematic sedimentological analysis of any of the rocks of the Moine Succession. Previous work with regard to the sedimentology of Moine rocks is thus largely confined either to descriptions of sedimentary structures at particular localities (Richey and Kennedy 1939; Wilson et.al. 1953; Soper 1960; Dalziel 1963; Powell 1964; Tobisch 1965), or to broad statements concerning the probable environments of deposition. Although there is broad agreement that a certain proportion of the Moine Succession accumulated under shallow-water conditions (Johnstone 1975), there is clearly a need for more precise environmental interpretations which would undoubtedly complement and clarify existing models for the evolution of the Moine.

### 7.1.2. Terminology

Sedimentological terminology follows accepted practice and it is consequently unnecessary to define many terms. 'Cross-stratification' refers to cross-bedded sets greater than 5 centimetres in thickness, and 'cross-lamination' to cross-bedded sets less than 5 centimetres in thickness. Within the latter category the inclined foresets may be of either wave-ripple or current-ripple origin. The terms 'bed' and 'laminae' refer to sedimentation units greater and less than 1 centimetres in thickness respectively. Throughout this chapter the metasediments are referred to in terms of their likely metamorphosed equivalents: thus quartzites and psammites are referred to as sandstones, and semi-pelites as siltstones.

### 7.1.3. Methods

At each locality the principal lithologies, thickness and geometry of bedding, together with the type and orientation of any sedimentary structures were all recorded. The differentation of primary sedimentary structures from tectonically-formed 'pseudo-sedimentary' structures may present a major problem for the geologist working in highly deformed sequences (Hobbs et al. 1976, Chapter 3; Whitten 1966, Chapter 13). Thus bedding, cross-stratification and rippled surfaces, for instance, all have their tectonically-induced counterparts. Detailed discussion with regard to the correct identification of certain primary sedimentary structures occurs at the appropriate points in this chapter. Extended logging is rarely possible except at exceptional localities; however, it is argued that it should still be feasible to demonstrate such gross features as cyclical sedimentation in areas of good exposure.

The metasediments have been subdivided into six separate lithofacies on the basis of lithology and sedimentary structures. The lithofacies concept forms a principal basis for sedimentological interpretation, and a facies is defined by Reading (1978, p.4) as ".... a body of rock with specified characteristics. In the case of sedimentary rocks it is defined on the basis of colour, bedding, composition, texture, fossils and sedimentary structures". In the present study the lithofacies have been grouped into five lithofacies associations. These are defined as ".... groups of facies that occur together and are considered to be genetically or environmentally related" (Reading op cit. p.5). Grouping in this manner is fundamental to environmental interpretation as the association provides additional evidence which makes interpretation easier than treating each facies in isolation.

## 7.2. LITHOFACIES DESCRIPTIONS

Each of the six lithofacies will be described and interpreted in detail.

### 7.2.1. Lithofacies 1: Siltstone

This common lithofacies occurs in several forms:

(a) As thin (less than 2 centimetres thick) beds and laminae, separating the sandy beds of lithofacies 2-6, and occasionally forming 'mud-drapes' which infill ripple forms. These thin beds tend to be laterally extensive and may often be traced for distances of up to 10 metres. Contacts with sandy beds tend to be gradational over several millimetres or may occasionally be sharp.

(b) As lenticular bodies within major stratigraphic units; these bodies may be up to 40 metres thick and 500 metres long.

(c) As a stratigraphic unit, the Druim Fearna Semi-Pelite, which is 250 metres thick and 3.5 kilometres long.

It seems likely that all occurrences of this lithofacies accumulated from suspension.

## 7.2.2. Lithofacies 2: Interbedded sandstone and siltstone

This lithofacies is distinguished by the regular alternation of layers of sandstone and siltstone. Where typically developed, thicker sandy beds (2-3 centimetres thick) alternate with rather thinner silty laminae and beds (0.5-2.0 centimetres thick) to impart a regular striped appearance. There is every variation in the scale and regularity of the striping: thus siltstone laminae may be as thin as 1-2 millimetres and separated by much thicker sandstone beds (Plate 60), or siltstone beds may exceed 10 centimetres in thickness and be somewhat thicker than adjacent sandstone beds (Plate 61).

Individual layers are usually planar and parallel, and within outcrop constraints it is often possible to trace them for distances of up to 10 metres. Occasionally, however, layers are less continuous and sandy layers may 'wedge out' (Plate 61), whilst silty layers may take the form of elongate lenticles (Plate 62). Layers may also be irregular in form (Plate 64).

Current or wave-formed sedimentary structures are rare within this lithofacies. Individual sandstone beds tend to be internally structureless (Plate 63) and preserve no sign of any sort of laminations. Possible cross-lamination is present at some localities (Plate 60) and undoubted current-ripples were observed at only one locality (Glen Suileag Banded Psammite: NN: 01808345), accompanied by small-scale graded bedding and slump structures.

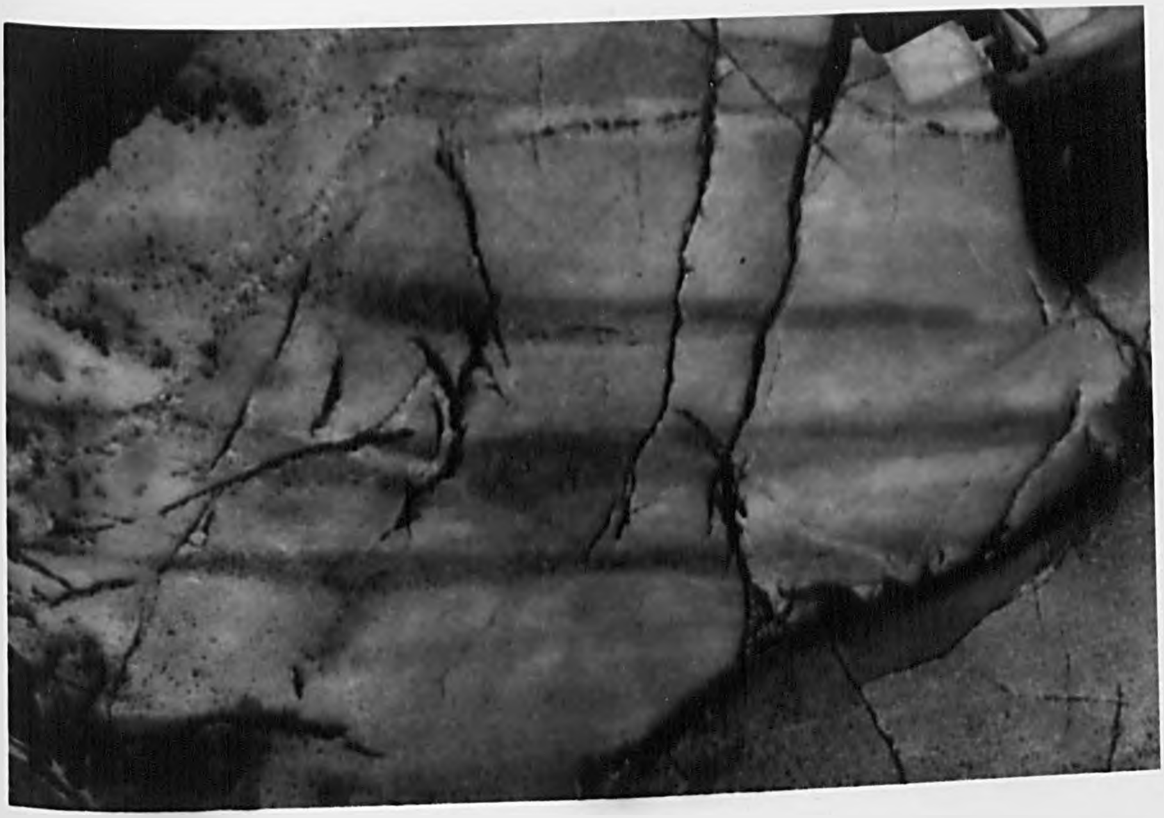
Interpretation. The distinctive nature of this lithofacies results from the rhythmic deposition of layers of sand and silt. Each individual layer was probably deposited as a discrete event:

Plate 60: Lithofacies 2: Sandstone beds separated by thin siltstone laminae. Possible cross-lamination at left. Sense of younging from left to right. (Kinlocheil Banded Quartzite, NM 960832).

Plate 61: Lithofacies 2: Relatively thick siltstone beds separate thinner sandstone layers in centre of photo. Sandstone layer in centre 'wedges' out from left to right. Outcrop inverted. (Kinlocheil Banded Quartzite, NM 963838).

Plate 62: Lithofacies 2: Lenticles of siltstone separated by sandstone layers. The silt appears to infill undulations in the sandy layers. (Kinlocheil Banded Quartzite, NM 98708435).

Plate 63: Lithofacies 2: Polished block demonstrating the internally structureless nature of the sandstone layers. (Stronchreggan Mixed Assemblage, NN 08557685).




0
6 cm.

Plate 64: Lithofacies 2: Interbedded sandstone and siltstone with thin lenticles and bands of calc-silicate. Irregular nature of layering in lower half is probably sedimentary in origin. (Glen Suileag Banded Psammite, NN 03258340).


0
6 cm.

it seems likely that the silt layers were deposited from suspension, with regular influxes of sand giving rise to the sandy layers. The precise conditions of deposition of these sandy layers is difficult to assess. They may also have accumulated from suspension, or possibly under low flow regime tractional flow conditions (Fig. 37 ) with a lack of suitable material inhibiting the formation of foreset laminae. Alternatively, they may have been deposited in the upper flow regime. The general lack of any current or wave-formed structures, channelling or erosive contacts through substantial thicknesses of sediment suggests that depositional conditions were relatively tranquil, and thus rapid deposition in the upper flow regime seems less likely than deposition from suspension or in the lower flow regime.

### 7.2.3. Lithofacies 3: Parallel-laminated sandstone

This lithofacies is the rarest of all six, and is difficult to identify except on clean washed stream outcrops. Accordingly, it may be more common than at present thought. It is characterised by sandstones which display even parallel laminations 3-4 millimetres thick. Occasionally these may attain thicknesses of 5 millimetres, and at certain localities it is possible to trace laminations laterally for distances of 1 metre.

Interpretation. Parallel-laminated sand may develop in several different ways. It is commonly due to the swash and backwash action of waves and is therefore commonly found on beaches or other sandy areas exposed to wave action. Such deposits usually display reverse

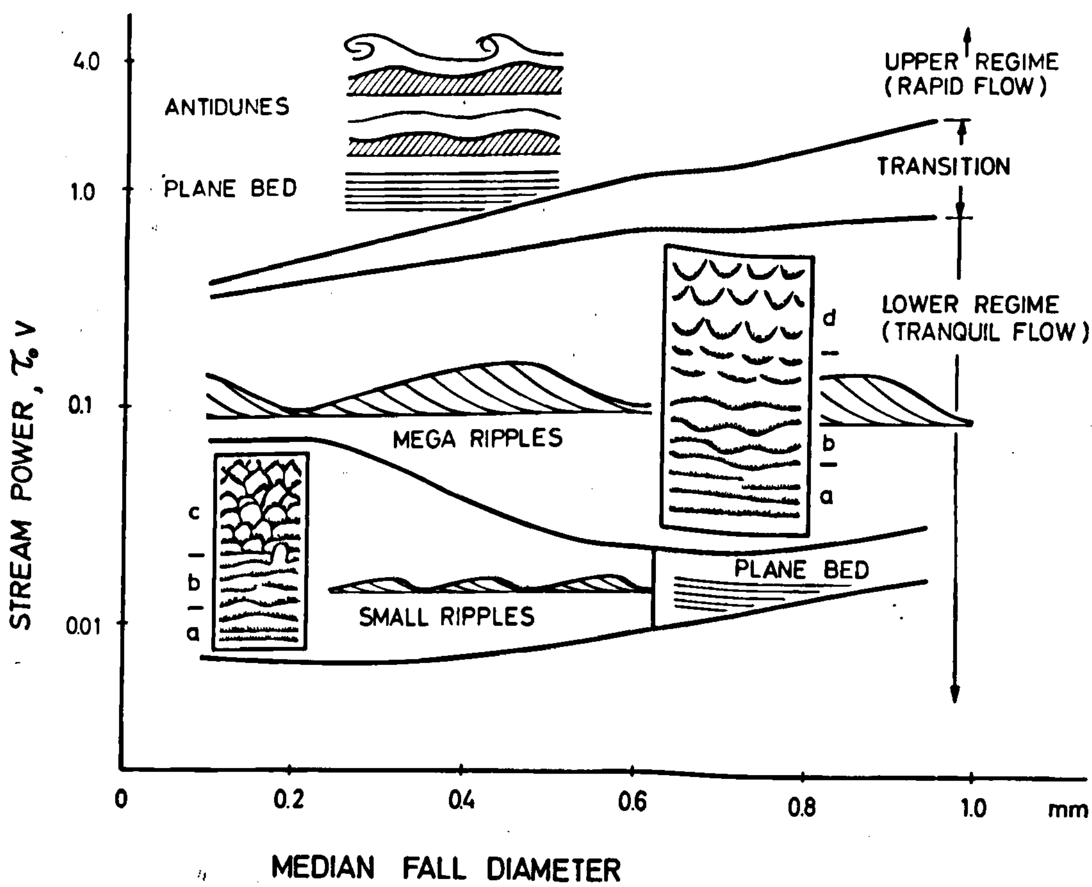
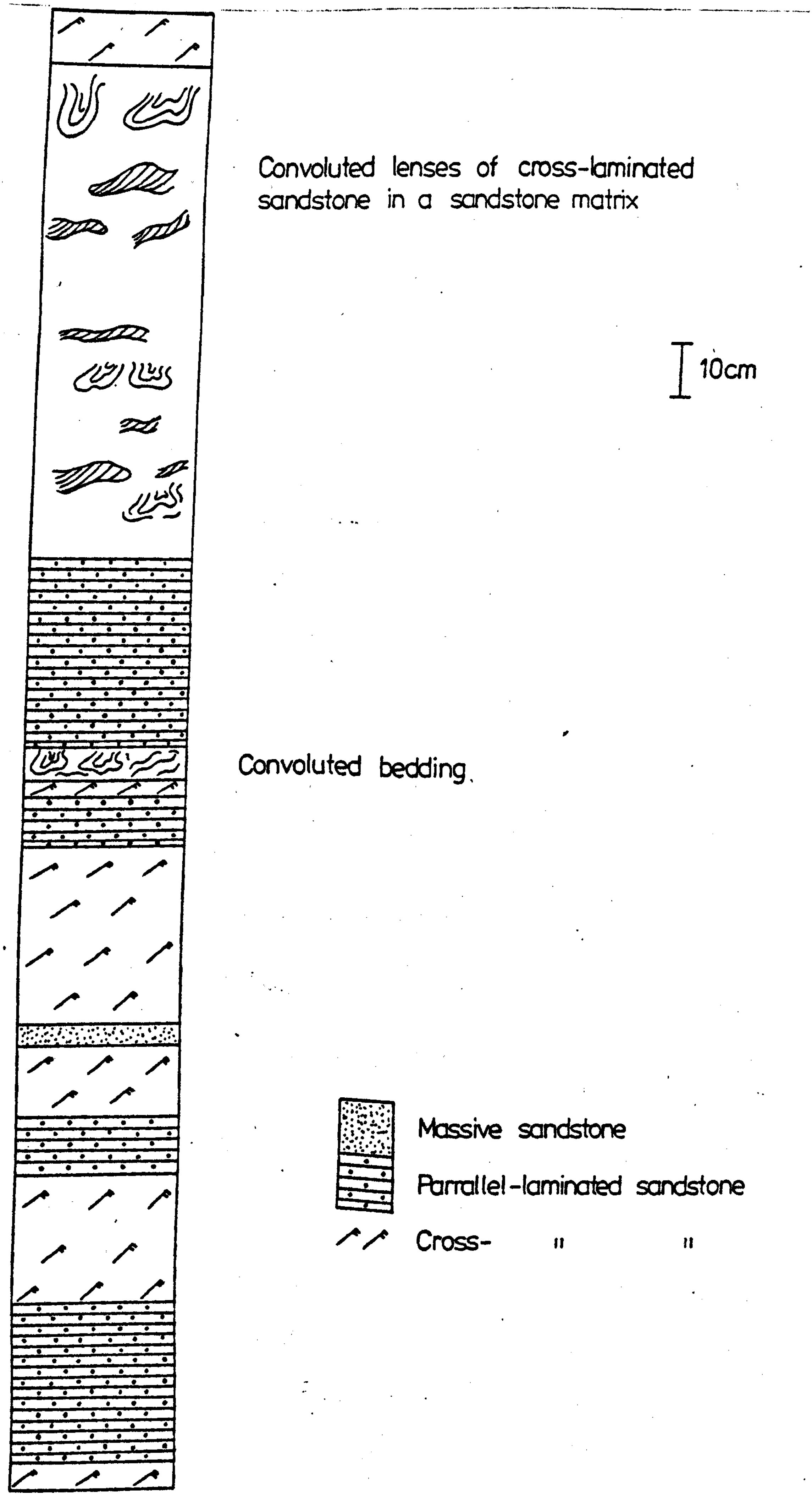



Figure 37: Schematic representation of various bedforms and their relationship to grain size and stream power. Taken from Reineck and Singh (1972, Fig. 2).

grading (Reineck and Singh 1973, p.104). It can also be produced in the plane-bed phase of the upper flow regime (Reineck and Singh op cit p.106). In this case the laminae are rather poorly-developed and of short lateral extent. A lower flat-bed exists within the lower flow regime for sand grains coarser than 0.6 millimetres, and for those coarser grain sizes an interpretation as lower flat bed deposits remains a valid alternative. Another important mode of genesis of parallel-laminated sand is through the sedimentation of suspension clouds at current velocities below those required for the genesis of ripples.

In Fig. 38 parallel-laminated sandstones up to 35 centimetres thick alternate with current-rippled sandstones of lithofacies 4. It is difficult to be certain about the flow regime under which the parallel-laminated sands were deposited, since original grain-size has been obscured. However, the close interbedding of the two lithofacies suggests broadly common conditions of deposition. The absence of any wave-formed cross-lamination or reversed grading makes it unlikely that wave action was responsible for this lithofacies. Similarly, the absence of any cross-stratification which might be interpreted as deposits of the upper part of the lower flow regime, suggests that deposition in the upper flow regime is an unlikely mechanism for the development of the sands under discussion. It therefore seems most likely that the parallel-laminated sands developed in response either to minor incursions of coarser sand, leading to the production of plane beds in the lower flow regime, or to the sedimentation of suspension clouds. In other cases there

Figure 38: Sedimentary log (Glen Garvan Psammite, NM 94807465).



is no close association with any lithofacies other than the massive sandstones of lithofacies 6, and it is accordingly more difficult to specify the mechanism by which the parallel-laminated sands were deposited.

### 7.2.4. Lithofacies 4: Cross-laminated sandstone

Cross-lamination is the most commonly observed type of sedimentary structure within the study area. A total of 239 individual sets were observed, of which 142 occurred in cosets. The sizes of sets are graphically represented in Fig. 39 . This lithofacies may be subdivided into current-rippled sandstone and wave-rippled sandstone.

#### 7.2.4.1. Current-rippled sandstone

Current-rippled sets generally differ only in size from morphologically similar cross-stratified sandstones of lithofacies 5. They occur both singly and in cosets, and are often interbedded with cross-stratified sets. Most current-rippled sets are tabular-bedded, although a rather greater proportion are trough-bedded (Plate 65) than in lithofacies 5. Both tabular and trough-bedded sets are associated at a bed-by-bed scale. The internal form of the foresets is generally simple: foresets are tangential to sigmoidal in form, and are defined by delicate laminae, 1 millimetre thick, and 2-3 millimetres apart. Tabular-bedded current-ripples may be traced laterally for distances of up to 3-4 metres. Occasionally ripples display undulatory tops which may be draped by siltstones of lithofacies 1, or immediately overlain by another current rippled set (Fig. 40 ).

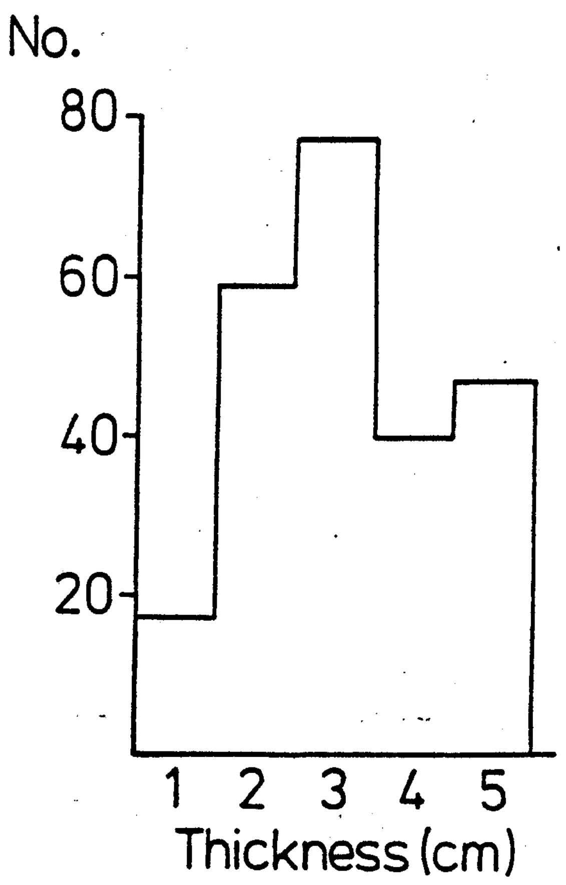



Figure 39: Thicknesses of cross-laminated sets.

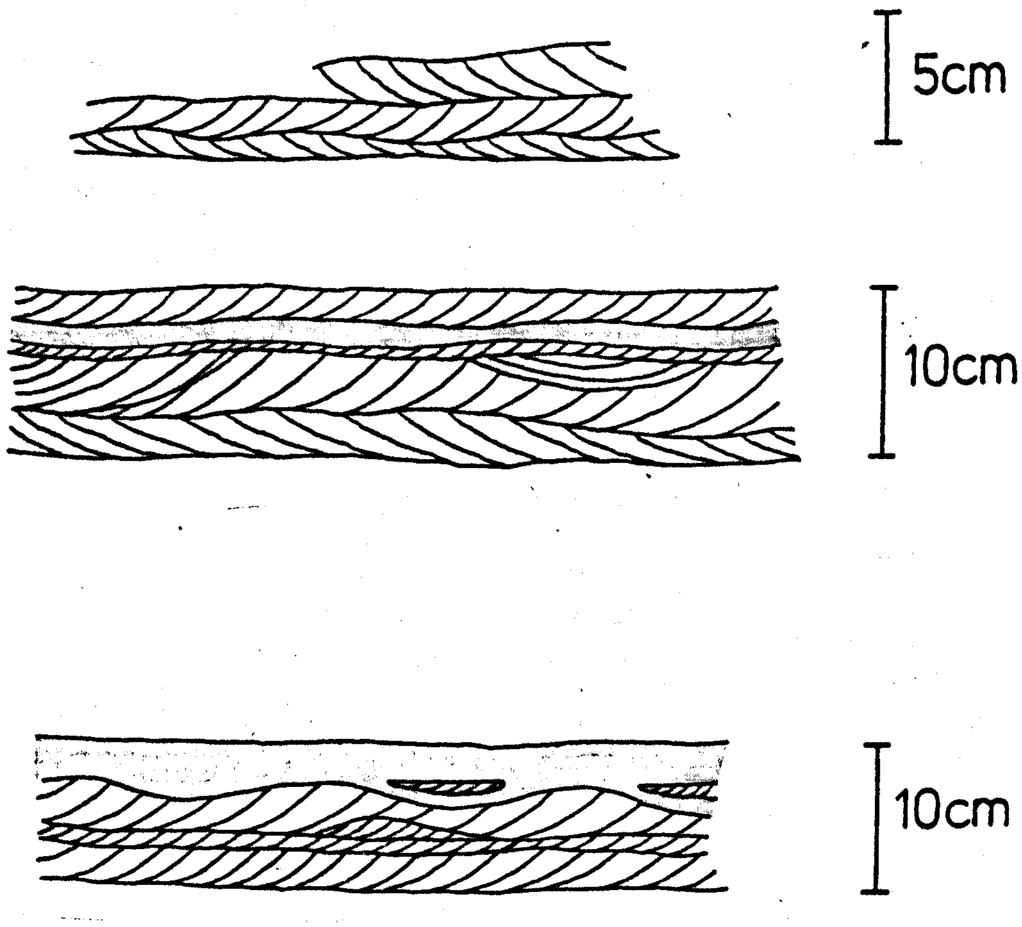
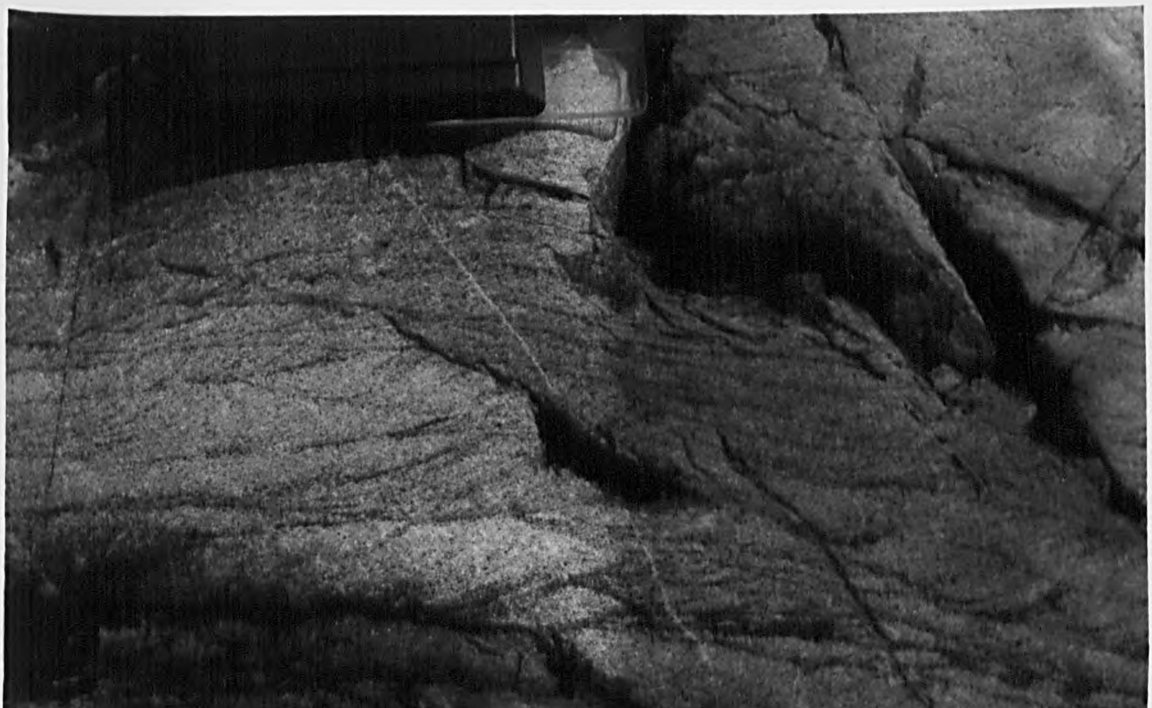



Figure 40: Current rippled sands with undulatory tops and silt drapes (Druim Fearna Semi-Pelite, NM 95307745).

Plate 65: Lithofacies 4: Trough-bedded current ripples.

(Glen Garvan Psammite, NM 94907465).



At 20 localities bidirectional "herring-bone" structure is present (Reineck and Singh op cit. p.86). In each case, current-rippled sets occur in cosets of 2-3 or more, and the foreset laminae in adjacent layers dip in opposite directions (Plates 66 and 67 ). At all these localities current-ripples are intimately associated with cross-stratified sets of lithofacies 5. Reineck and Singh (op cit p.86) emphasise that the correct identification of this structure depends upon the availability of three-dimensional sections: "even if the mean difference in the direction of dip of foreset laminae in adjacent layers is less than $90^{o}$, in certain sections they are seen to be oppositely-dipping; festoon-bedding seen in diagonal sections may look like herring-bone cross-bedding". At several localities it is possible to recognise the structure unequivocally by means of three-dimensional sections. At the remainder, where this is not possible, the current-bedded sets are tabular in form and the angle between the bedding and foreset laminae is, in each case, between $25^{o}$-$30^{o}$. These angles are close to the maximum angle of repose of $35^{o}$ in undeformed sediment (Reineck and Singh op cit p.15) and this implies that the sections being viewed are approximately normal to the direction of current-flow. Thus true herring-bone structure is inferred to be present at these localities also.

'Dish' structures (Wentworth 1967) are present within current-rippled sandstone at localities in the North Garvan River (e.g. NM 94957470). They appear as gently concave-up arcs, with amplitudes of 1-2 centimetres and wavelengths of 3-4 centimetres (Plate 68). Occasionally they may form anastomosing patterns up to 13 centimetres thick (Plate 69). They are morphologically similar to those described

Plates 66 and 67: Lithofacies 4: Tabular and trough-bedded current ripples, displaying bi-directional "herring-bone" structure. (Cona Glen Psammite, NN 035784).

Plate 68: Lithofacies 4: Minor development of 'dish' structure. (Glen Garvan Psammite, NM 94907465).

Plate 69: Lithofacies 4: More extensive development of 'dish' structure as a laterally extensive band 13 centimetres thick. (Glen Garvan Psammite, NM 94907465).

by Stanley (1974) and Walker (1979), and are readily distinguishable from cross-lamination of wave-ripple origin, the diagnostic features of which are outlined by Dr Raaf et al (1977).

Units of convoluted lamination up to 90 centimetres thick occur within current-rippled sandstones in the North Garvan River (Figs. 38 and 41 ). They are characterised by chaotic assemblages of disoriented detached lenses of current-rippled sands "floating" in a silty matrix. Some may be elongate and sub-parallel to bed boundaries, whilst some may be tightly folded into a concave-upwards form. Plate 70 shows the chaotic folding of alternating sand and silt layers, although individual layers are still traceable laterally for some distance. The sense of vergence associated with these demonstrably 'soft-sediment' folds is in the same direction as the current flow within the sandstones immediately below the convoluted horizon.

7.2.4.2. Wave rippled sandstone

Localities where wave-ripples are present in cross-sectional profile are few in number, but they are well-developed on Druim Fearna (NM 95307745). Here, the wave-ripples are closely associated with current-ripples in thin sandstone beds which have parallel sides and sharp contacts with the enclosing silts and muds of lithofacies 1. The internal structure of these ripples is illustrated in Fig. 42 and shows several of the features which are generally thought to be diagnostic of wave-ripples (Fig. 43 ), particularly the gradation of parallel lamination into cross-lamination, and the presence of bundled bi-directional lenses and intricately interwoven cross-lamination. Occasionally, complexly interwoven concave laminations

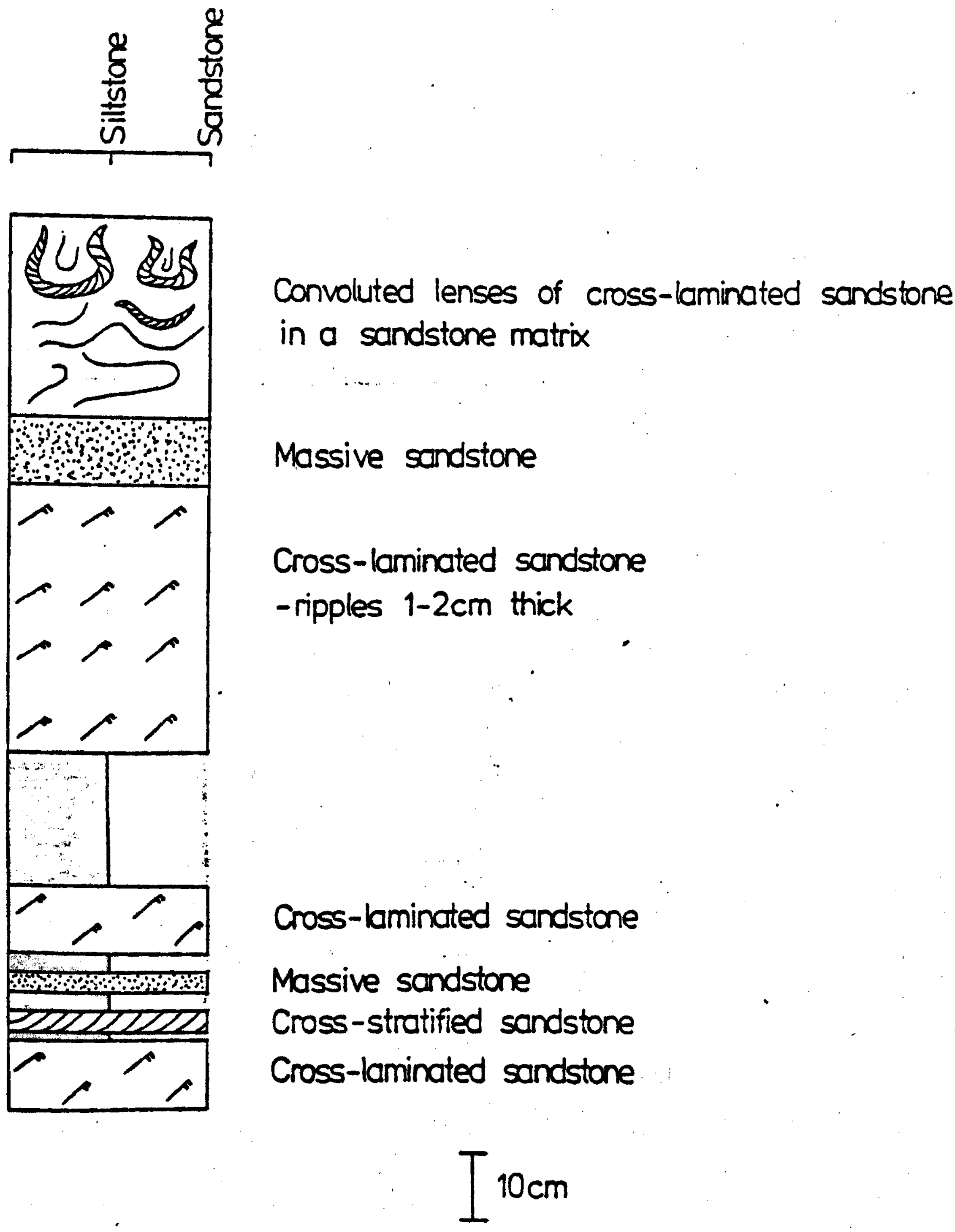



Figure 41: Sedimentary log (Glen Garvan Psammite, NM 945743).

Plate 70: Lithofacies 4: Syn-sedimentary folds overlying tabular and trough-bedded current ripples. (Glen Garvan Psammite, NM 94857465).

Plate 71: Cross-stratified sandstone of lithofacies 5, bounded by siltstones of lithofacies 1. Sense of younging from right to left (Kinlocheil Cross-Bedded Quartzite, NM 923729).

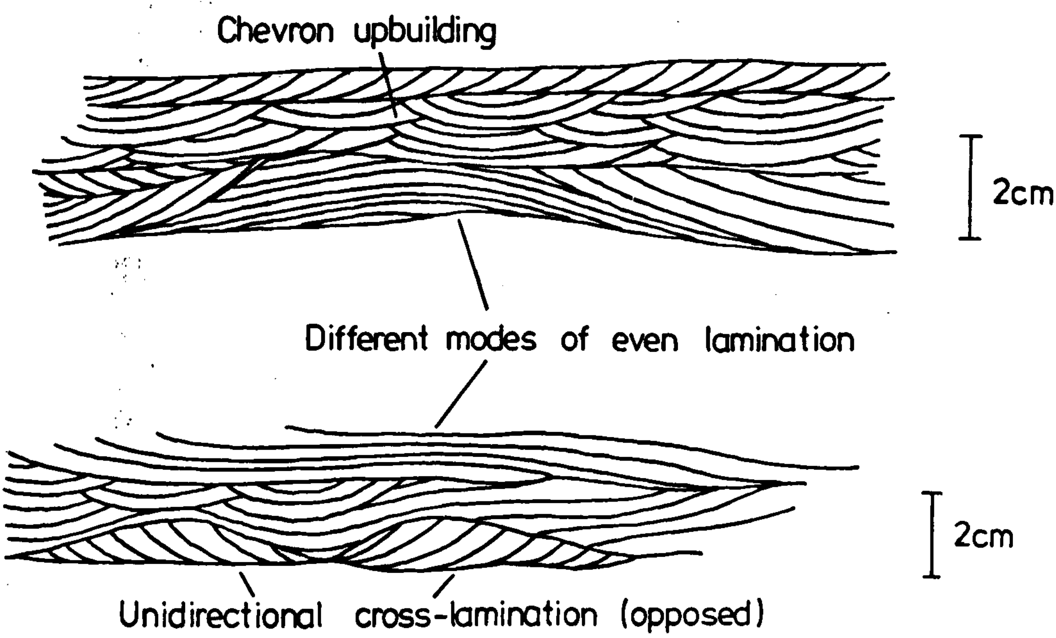



Figure 42: Wave-rippled sands (Druim Fearna Semi-Pelite, NM 95307745).

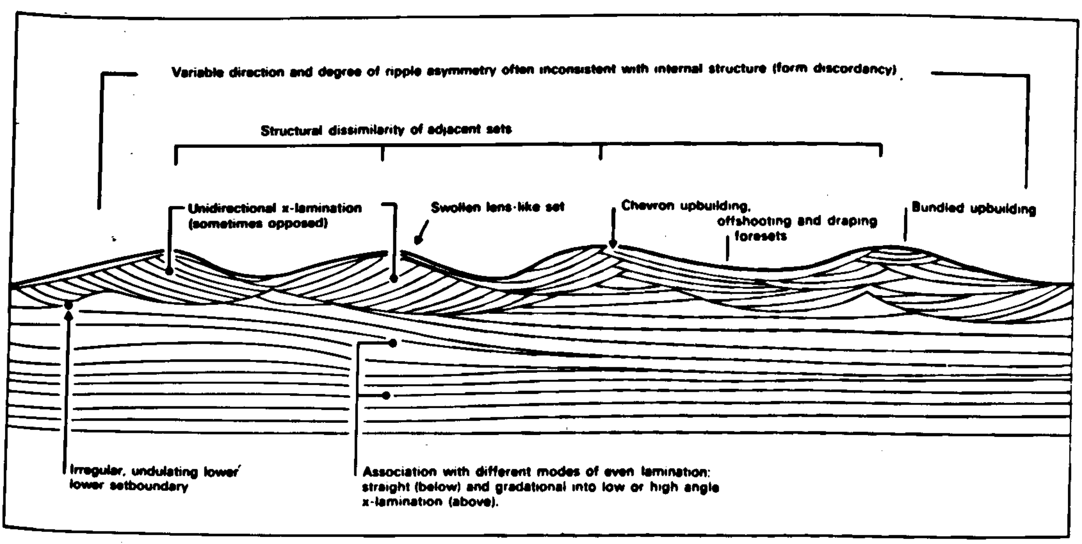



Figure 43: Diagnostic features of wave-ripples (compare with above). Taken from Johnson (1978, Fig. 9, 47).

may indicate the presence of wave-ripples at other localities. However, identification is less certain in these cases, mainly due to the state of weathering at the exposures in question.

Interpretation. The formation of lithofacies 4 may be attributed primarily to the migration of small-scale ripples under conditions of the lower flow regime (Fig. 37), and secondarily to the subsequent reworking of the sands by wave action. Cosets of ripples form under conditions of net sedimentation. The presence of bidirectional current-formed structures is particularly important, indicating regular reversals in the direction of tractional currents. This feature has long been considered to be indicative of net deposition by tidal currents, ripples migrating in opposite directions in the ebb and flood stages of the tidal cycle.

'Dish' structures are indicative of abundant fluid escape during the deposition of the sandstone, and imply rapid deposition of a sand bed from a fluidised flow (Walker op cit p. 94). However, the syn--sedimentary folds illustrated in Plate 70 and the convoluted lamination probably originated in a different manner. The observation that the sense of vergence associated with the syn-sedimentary folds is in the same direction as the direction of current flow adjacent to the folded layer, and elsewhere in the North Garvan River, suggests that the folding was accomplished by drag during the deposition of the overlying bed. Rettger (1935) showed that this was only possible when the deformed bed was liquefied at the time of the deposition of the overlying bed. The work of Seed (1968) showed that a cohesionless sand will liquefy when it is subjected to large cyclical stresses such as those occurring during earthquakes. Following their work, Allen & Banks (1972) argue that the only plausible manner by which current drag may cause syn-sedimentary folding is by action on a bed already liquefied as a result of earthquake shock. The convoluted lamination provides further evidence that indicates that the cross-laminated sandstones were deposited during a period of seismic activity. Johnson (1977) also regards liquefaction following an earthquake shock as a primary cause of these structures. The temporary increase in pore-fluid pressure during the shock renders the bed quasi-liquid, and differential deformation and the formation of convolutions occurs as pore-fluids are displaced upward through the bed, resulting in the re-establishment of a grain-supported fabric. Since such deformation affects single units only, it is likely that the process of deformation occurred at the sediment-water interface.

### 7.2.5. Lithofacies 5: Cross-stratified sandstone

Examples of cross-stratification were observed at 75 localities, and 127 individual cross-stratified sets were recorded, of which 63 occurred in cosets. The sizes of the cross-stratified sets are graphically represented in Fig. 44.

Cross-stratified sets are almost entirely tabular in form, with parallel sides and non-erosive bases (Plate 71). Only rarely are sets observed to "wedge-out" laterally (Plate 72); many may be traced laterally for distances of 4-5 metres within the constraints of outcrop size. Foreset laminae are generally angular to tangential in form, but may occasionally be sigmoidal. Individual laminae are c.1 millimetre thick and 3-4 millimetres apart. Examples of re-activation surfaces and reversely-dipping sets are present at one locality (Fig. 45).

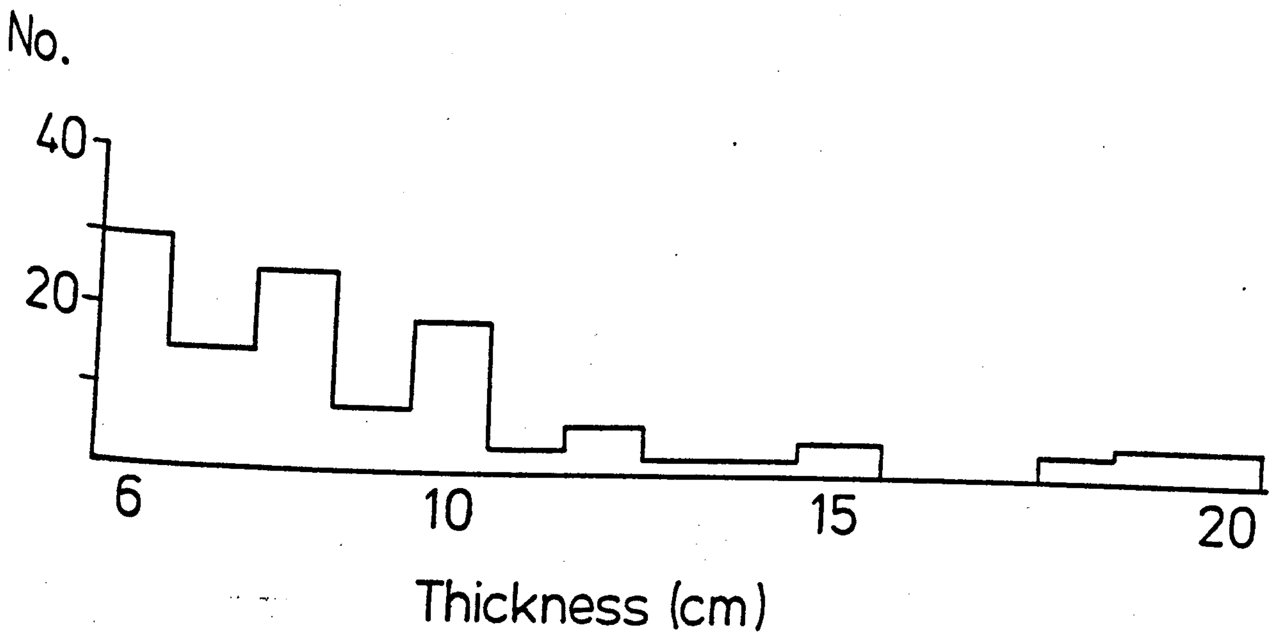



Fig. 44: Thicknesses of cross-stratified sets.

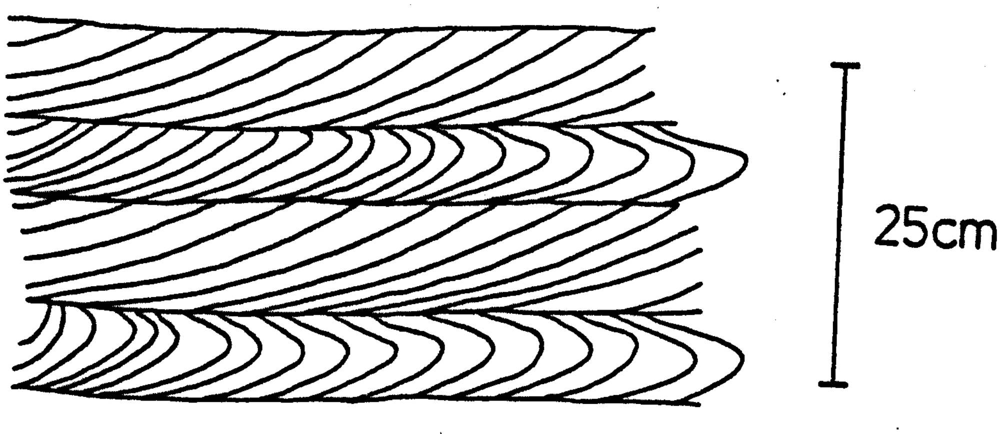



Figure 46: Syn-sedimentary folds (Basal Psammite, NM 91657310).

Plate 72: Cross-stratified sandstone of lithofacies 5. Lower two sets are tabular in form with planar to sigmoidal foresets. Upper set 'wedges' out laterally left to right. (Kinlocheil Cross-Bedded Quartzite, NM 96057750).

Plate 73: Lithofacies 5: Asymmetrical syn-sedimentary fold. (Kinlocheil Cross-Bedded Quartzite, NM 92457290).

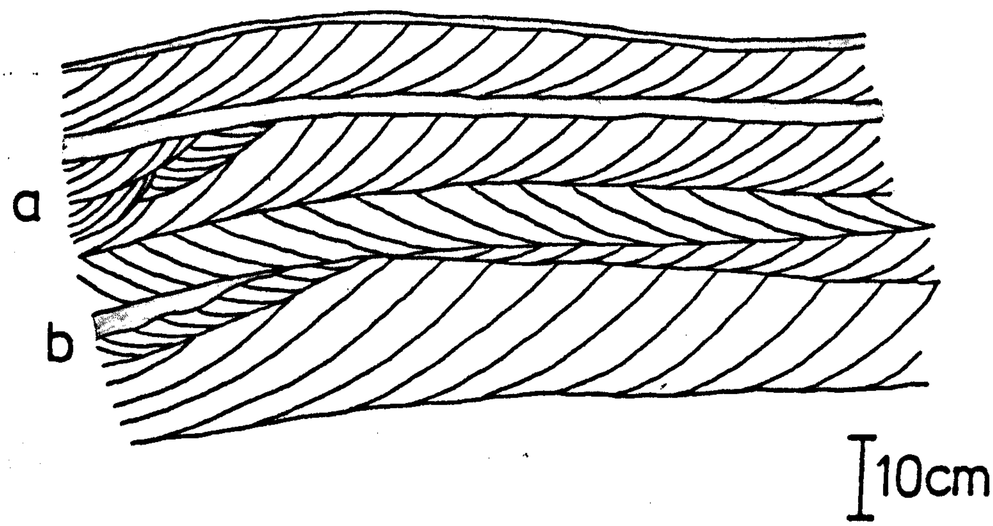



Figure 45: Reactivation surfaces (a) and reversely-dipping sets (b). (Cona Glen Psammite, NN 0357840).

At 8 localities cross-stratified sets form bidirectional "herring-bone" structure, in association with the cross-laminated sands of lithofacies 4. Once more, the availability of three-dimensional sections at several localities enables the correct identification of this structure, while at other exposures the relatively steep angles between foresets and bedding suggests that the sections under view are approximately normal to the direction of current flow.

Structures which could be either deformed and overturned cross-stratified sets, or tectonic folds, are exposed on water-polished surfaces in the Cona River within the Kinlocheil Cross-Bedded Quartzite (NM: 92357290). A total of 4 individual folds were observed, none of which were adjacent, however. Most are asymmetrical (Plate 73), although some approach a symmetrical form (Plate 74). None display any internal variation in style along the length of the bed. It is clearly essential to decide whether these structures result from primary deformation during sedimentation, or from subsequent tectonic deformation. The following observations are pertinent with regard to this:

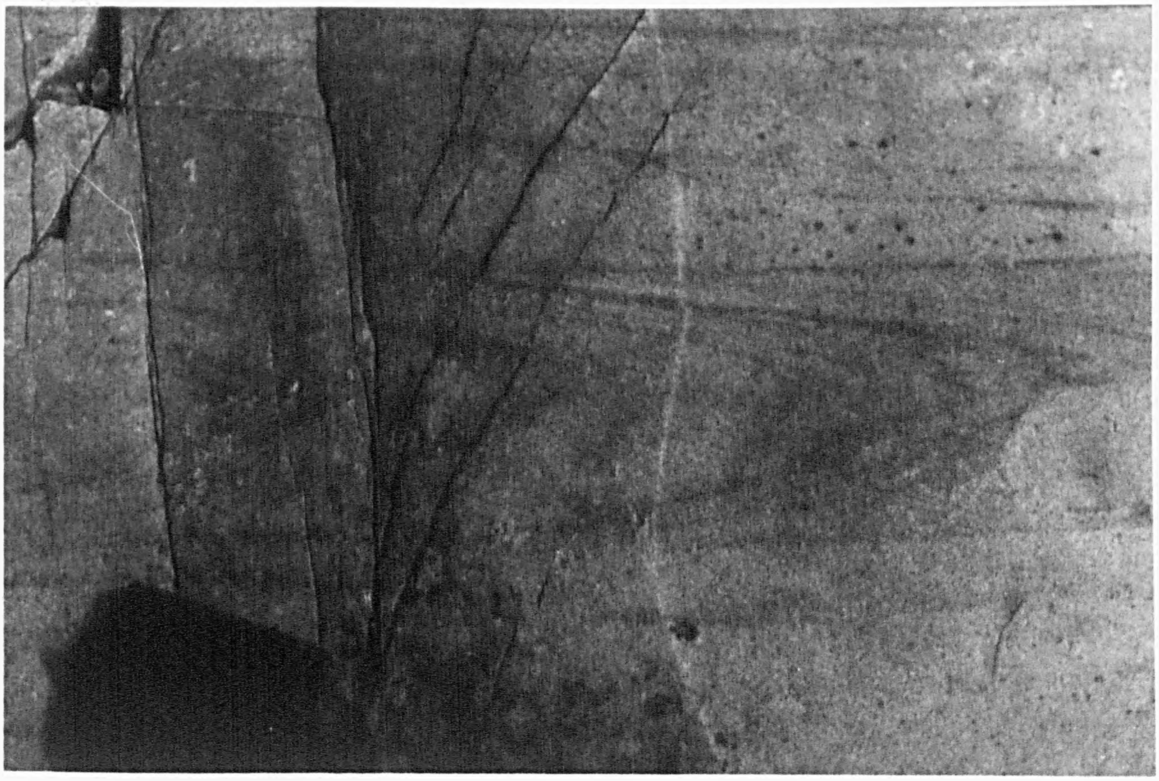
(a) No fold displays any internal lithological variation along the length of the bed.

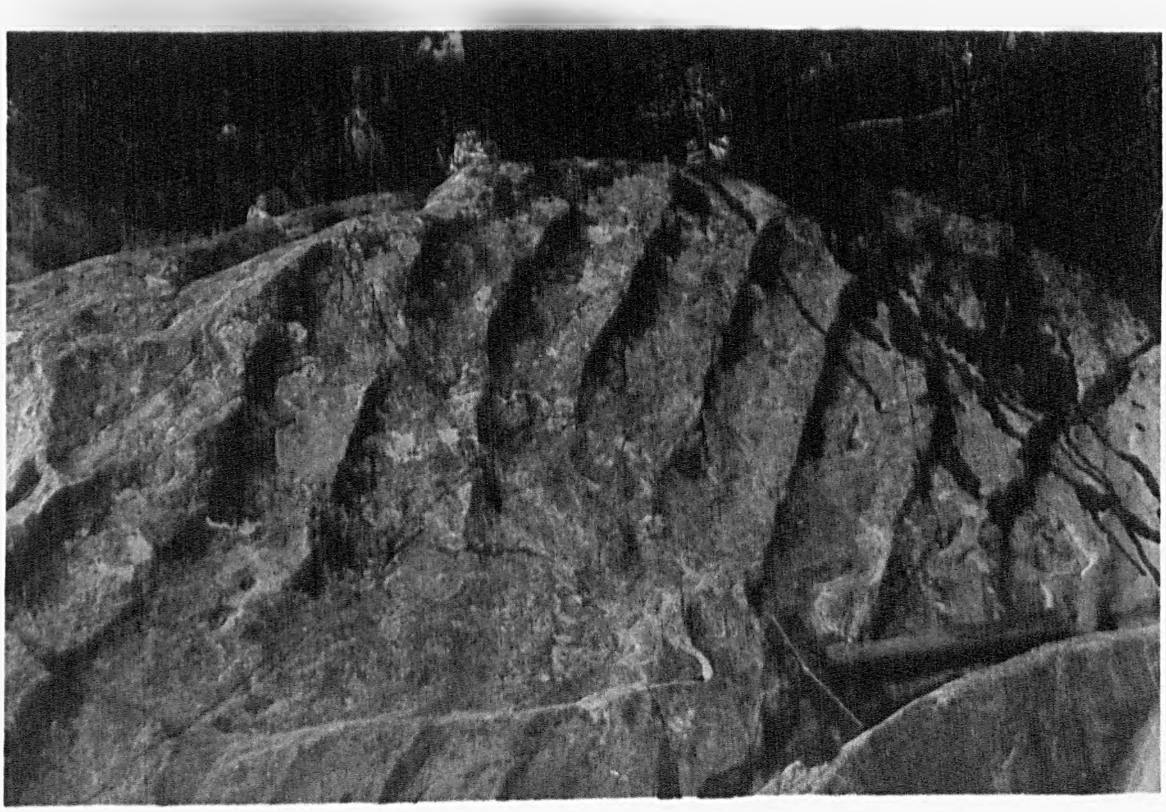
(b) Most are composed of equally spaced laminae identical to those of undoubted cross-stratified sets.

(c) All show the same southerly sense of closure. Most cross-stratified sets within the Kinlocheil Cross-Bedded Quartzite show a northerly direction of sediment transport (p. 171), and if any of these were overturned, a southerly-closing fold would result.

Plate 74: Lithofacies 5: Symmetrical syn-sedimentary fold. (Kinlocheil Cross-Bedded Quartzite, NM 92457290).

Plate 75: Lithofacies 6: Symmetrical wave-ripples preserved on the S0/S1 surface. (Glen Garvan Psammite, NM 957754).





These observations strongly suggest that these features are deformed cross-stratified sets and not tectonic folds. Moreover, it seems likely that the deformation was approximately contemporaneous with deposition, and not tectonic, since this would require small-scale sinistral shearing for which there is no other evidence.

However, clear evidence for syn-sedimentary deformation of cross-stratified sets occurs at a locality adjacent to the Cona River (NM 91657310) within the Basal Psammite. Here, there is a coset of 4 individual sets, with 2 sets displaying great variation in style along the length of the bed, from a slight overturning to a very nearly isoclinal form (Fig. 46). These folds are very similar to those described by Tobisch (1965) from the Cannich area. The observation that the most deformed sets are separated by relatively undeformed sets implies that the deformation of any particular unit took place before or during the deposition of the set immediately overlying it, and hence the deformation is essentially primary.

Interpretation. The formation of tabular sets of cross-strata is commonly attributed to the migration of sand waves under conditions of the lower flow regime (Harms et al. 1975). Where the foresets depart in form from angular or tangential to sigmoidal, this may be taken as an indication either of increasing current velocity, and hence suspension load, or of a change in particle size (Reineck & Singh 1973, p.21). As metamorphism has obscured original grain-size differences, it is difficult to assess the precise reasons for this. Occasional trough-bedded sets represent the migration of dunes, and the change in bedform from sand waves may be attributed to a temporary increase in energy (Harms et al. op cit p.48). Cosets form under conditions of net sedimentation. The presence of bidirectional current-formed structures is again attributed to deposition by tidal currents. It is important to note that the only examples of reactivation surfaces and reversely-dipping sets are recorded in association with "herring-bone" structures. These structures resemble those recorded from tidal sand waves in the Lower cretaceous Woburn Sands (Bentley 1970; De Raaf and Boersma 1971), although those observed within the Loch Eil Division are on a much smaller scale. The thicker sets are interpreted as sand waves which record the direction of the strongest tidal currents; however, the reversing tidal currents were sufficiently powerful to modify the sand wave lee slopes, and to generate up-slope migrating dunes (Klein 1970). As reactivation surfaces may form also in the fluvial environment as a product of fluctuations in discharge (Collinson 1970), this structure cannot by itself be used to identify tidal activity, but when observed in association with bi-directional current formed structures it may provide additional evidence of tidal activity.

The deformed cross-stratification described are of the "Type (a)" of Allen and Banks (1972), where the overturning of the foresets forms a single recumbent fold whose axial plane is approximately parallel with the base and top of the deformed unit. Allen and Banks attribute such structures to the deformation of a liquefield sand by current drag following an event such as a seismic shock. This is thus identical to the method of formation of the rather more chaotic syn-sedimentary folds illustrated in Plate 70, the origin of which has already been discussed (p. 152).

### 7.2.6. Lithofacies 6: Massive sandstone

This lithofacies is volumetrically the most important, and is characterised by thinly- to thickly-bedded sandstones which generally present a homogeneous aspect. Individual beds are parallel-sided and may be traced for lateral distances of up to 10 metres within the constraints of outcrop size. They were only observed to "wedge-out" at one locality, in the An-t-Suileag (NN 01958110), and their bases and tops are usually planar and non-erosive. A poorly-developed 'flat' lamination may occasionally be detected within these sands, but internal structures are rare, even on clean washed stream exposures. A large number of blocks have therefore been sliced in an attempt to obtain clean surfaces for examination. Most of these samples did not reveal any further information relating to the internal structure of the sands, although a few revealed hitherto unsuspected cross-lamination and parallel lamination.

The massive sands may occasionally show grading from sandy bases to silty tops. Examples of such grading occur on the NE slopes of Druim Beag where at one locality (NN 01408275) five 50 centimetre thick graded beds occur in succession. Similar grading is also sporadically present in Banavie Quarry (NN 11207835). 'Dish' structures were observed within massive sands on Glas Bheinn (NM 93857590); they are morphologically identical to those described from the North Garvan River (p. 149 ). The tops of massive sands display rippled surfaces at three localities. At Fassfern (NN 02107895) massive sands display symmetrical ripples on their upper bedding surfaces; at least 4 individual surfaces are preserved. The ripples have sharp crests which are 8 centimetres apart and are separated by rounded concave troughs. The ripple crests were not observed to bifurcate. The amplitudes of the ripples are generally uniform at 1 centimetre. These rippled surfaces are morphologically identical to a similar surface at a locality in the North Garvan River (Plate 75). At this locality ripple crests are approximately 14 centimetres apart. Rather different morphology is shown by a rippled surface at (NN 03557840), 1 kilometre E of Fassfern. Here, the ripples are symmetrical, but have rounded crusts and pointed troughs,and the crusts show a high degree of bifurcation and amalgamation. These examples of rippled surfaces are identical to those observed both in contemporary sedimentary environments (Reineck & Singh 1973, Chapter 2), and in ancient sediments (Singh 1968). Furthermore, they bear no relation to "pseudo-rippled" surfaces which are commonly developed in lithofacies 1 due to the intersection of cross-cutting crenulation fabrics with bedding-parallel fabrics.

Interpretation. The absence of traction-formed sedimentary structures means that it is somewhat more difficult to specify the precise process by which these sands accumulated. However, it is clearly imperative to do so, as this lithofacies constitutes a substantial proportion of the succession. It is possible that some of the more thinly-bedded sands of this lithofacies may have been deposited from suspension, but this is an unlikely mechanism to account for the thickness of sediment in question. Two alternative models may be proposed for the development of thick sequences of generally massive and structureless sands:

(a) The first model envisages these laterally extensive tabular sands to be the product of the migration of sand mass under essentially the same flow regime as the cross-stratified sands of lithofacies 5. However, in this case, the lack of traction-formed sedimentary structures may be attributed to lack of material suitable for the delineation of foresets. This is particularly common in mature well-sorted sands which are clay poor, possibly as a result of a prolonged history of reworking. Thus Klein (1977, p.110) comments: ".... most quartz arenites show a massive and structureless appearance on outcrop ....", with reference to tidally-deposited sands. An example of this is recorded by Klein (1970) from the Lower Fine-Grained Quartzite (Middle Dalradian) of Islay, which includes a proportion of massively-bedded sands which have also been interpreted by him to result from deposition by tidal currents in a shallow marine subtidal environment.

(b) An alternative model attributes the lithofacies to the very rapid deposition of high concentrations of sediment, which consequently inhibits the development of tractional features. There are four basic mechanisms by which this might be achieved:

(i) Grain Flows. These are concentrated dispersions of cohesionless sediment, which move downslope in response to the pull of gravity, the dispersion being maintained by intergranular collision. They form on the leeside of bed forms and deltas, and invariably involve relatively small volumes of sediment, usually forming massive beds less than 5 centimetres thick. The only conditions under which grain flows form very thick beds is where they transport gravel-sized debris on low slopes (Middleton 1970; Lowe 1975). The sands under discussion do not contain gravel-sized debris, and beds usually exceed 5 centimetres in thickness, and consequently it seems unlikely that grain flow operated as a significant mechanism of sand transport.

(ii) Debris Flows. These are downslope movements of heterogeneous mixtures, including clay and water in response to gravity. They may often include a high proportion of lithic fragments and tend to be poorly sorted. Both these features contrast strongly with the massive sands under discussion.

(iii) Liquefied Flows. Coarse sediment liquefied after slumping or by compaction of underlying unconsolidated clay layers can flow down slopes of only a few degrees. (Lowe 1976). Grains settle through the liquid, displacing it upward. Pre-existing tractional sedimentary structures are thus destroyed and are replaced by de-watering structures such as fluid-escape pipes with sand volcanoes, dish structures, internal load-structures, and convolution-like folds. These structures are, however, generally absent in the Loch Eil Division rocks and this suggests that a liquefied phase was not normally developed within the massive sands.

(iv) Turbidity Currents. The major features of "classical turbidites" are listed by Walker (1979), p.91). They tend to have abrupt, sharp bases and grade upward into finer sand, silt and mud. Tool and scour marks are commonly found on their undersurfaces. Furthermore, a large number of turbidite sequences wholly or partially preserve, on a repeated bed-by-bed scale, a sequence of sedimentary structures known as a 'Bouma sequence', indicative of a progressively waning flow regime. The massive sands in question may often have sharp bases, but these are not developed in association with tool or scour marks, and in any case are no more frequently encountered than gradational contacts with underlying thin beds of silt and mud. Internal grading is rarely present, and no undoubted Bouma sequences were identified. It therefore seems unlikely that turbidity currents were important contributors to the accumulation of the massive sands.

In conclusion it seems most likely that the massive sands of lithofacies 6 are the result of the deposition of clean well-sorted sands as laterally-extensive sand waves under conditions of the lower flow regime, rather than representing the products of the repeated rapid deposition of high concentrations of sediment. This interpretation is broadly supported by the interbedding of massive sandstone with cross-laminated and cross-stratified sandstone (Fig. 49 ), which suggests a common origin. Given that a certain amount of sedimentological detail tends to be obscured on worn hillside exposures, it is entirely possible that a proportion of this lithofacies may be more correctly assigned to one or more of lithofacies 3, 4 or 5. The presence of occasional grading implies deposition from waning currents. A degree of local liquefaction is suggested by the presence of "dish" structures on Glas Bheinn. Symmetrical ripple forms are diagnostic of reworking by wave activity (Reineck and Singh op cit, Chapter 2). The origin of the flat and rounded tops of the rippled surface at (NN 03557840) may reflect planing-off during tidal reversal (Klein 1970), or a more extensive erosional history over a period of several tidal cycles. In this context it is significant that bidirectional current-formed cross-lamination and cross-stratification are well-developed at the same locality.

## 7.3. LITHOFACIES ASSOCIATIONS

The six lithofacies which have been described, may be grouped into five major lithofacies associations. Following the definition of Reading (1978 p.5) each lithofacies contains groups of facies which occur together and are considered to be environmentally related. This concept is basic to all environmental interpretation when combined with Walther's Law of Facies, which indicates that facies occurring in a conformable vertical sequence were formed in laterally adjacent environments. As the metasediments have already been stratigraphically subdivided purely on the basis of lithology, it is inevitable that the different stratigraphic formations described in Chapter 2 will closely correspond to both the different lithofacies and lithofacies associations. The five lithofacies associations have been designated A, B, C, D and E.

### 7.3.1. Association A

This is characterised by the predominance of the siltstones of lithofacies 1, in association with thinly-bedded sandstones. These sandy beds have parallel sides and sharp contacts with the enclosing siltstones and are often traceable for distances of up to 20 metres. They may be internally massive, or preserve cross-lamination of both current and wave origin. Stratigraphically the association corresponds to the Druim Fearna Semi-Pelite. Fig. 47 presents a vertical section of a part of this association, which contains members of lithofacies 1, 4 and 6.

### 7.3.2. Association B

This differs from Association A only in the greater proportion of sandstone. Thus the representative section through this association (Fig. 48 ) depicts a sequence in which beds of sandstone and siltstone alternate, and are present in approximately equal proportions. The sandy beds are similar to those of association A in that they are

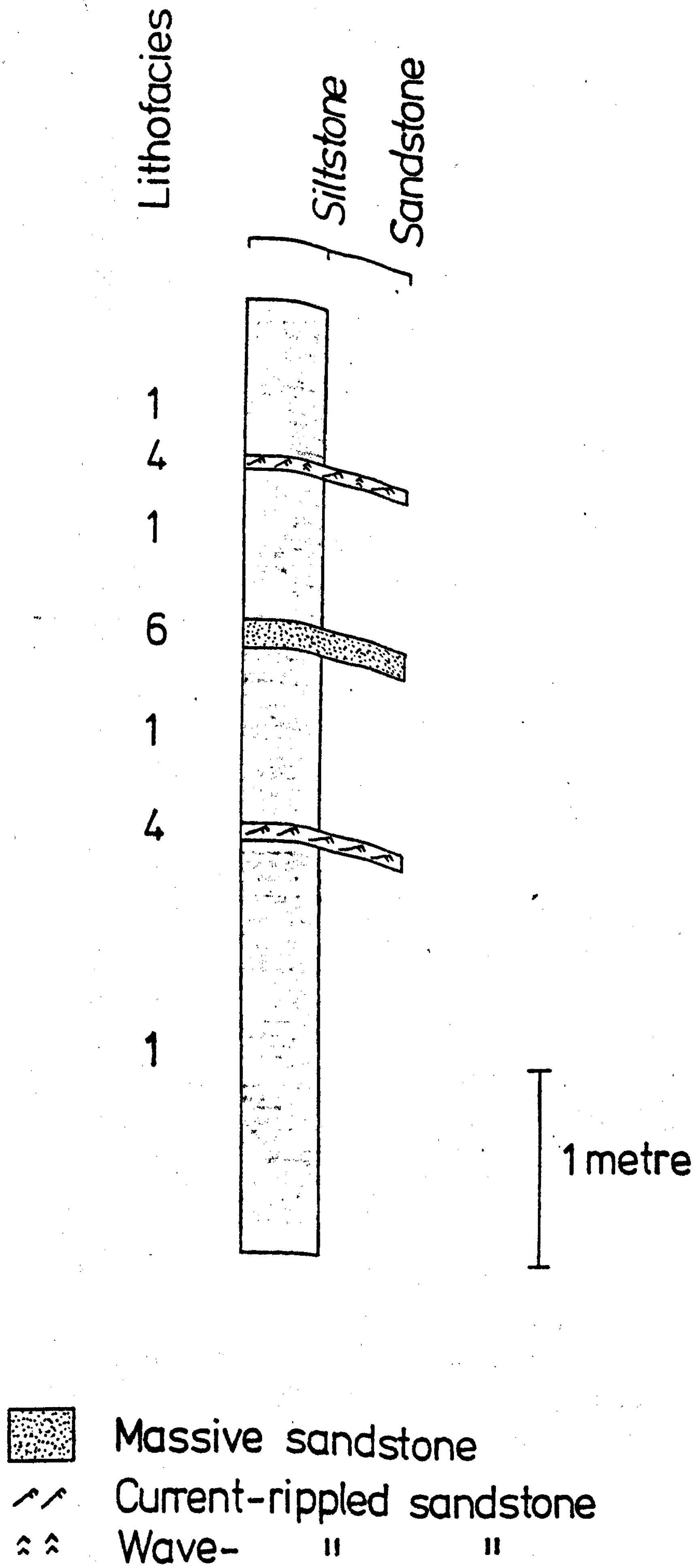



Figure 47: Representative log of Association A (Druim Fearna Semi-Pelite, NM 95307745).
Interpretation: Dominantly deposition of silt from suspension. Occasional influxes of sand, which were also probably deposited from suspension, and subsequently reworked by current and wave processes.

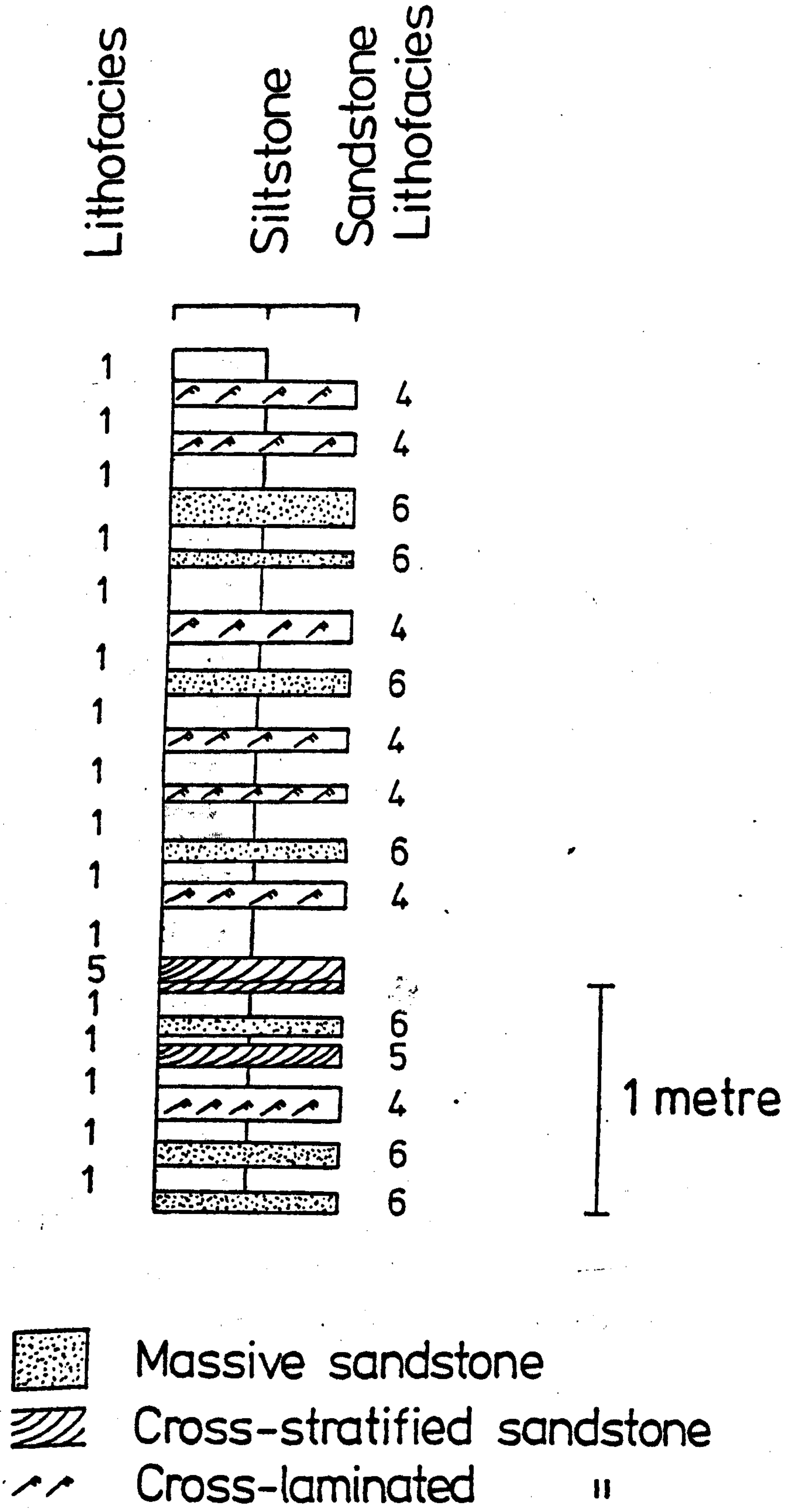



Figure 48: Representative log of Association B (Druim Fearna Semi-Pelite NM 957776 ).
Interpretation: Deposition of silt from suspension alternating with influxes of sand. Some sand also accumulated from suspension, but some sand layers resulted from the migration of laterally extensive sand waves.

parallel-sided and laterally extensive. They commonly preserve cross-lamination of current origin, which may be arranged in cosets and form 'herring-bone' structure. A few sandy beds are internally massive. This association forms a transitional sequence between the Druim Fearna Semi-Pelite and the Kinlocheil Cross-Bedded Quartzite.

### 7.3.3. Association C

This comprises elements of all the lithofacies described; it is, however, characterised by the predominance of sandy beds, mainly of lithofacies 6. Cross-laminated and cross-stratified sandstones occur at intervals, and may form 'herring-bone' structure. Beds of lithofacies 2 and 3 may occur, but are generally rare. Siltstones are common within this association, generally forming thin beds, 1-2 centimetres thick, which separate sandy beds. This association corresponds stratigraphically to the:

Basal Psammite
Kinlocheil Cross-Bedded Quartzite
Glen Garvan Psammite
Druim Fada Quartzite
Cona Glen Psammite

Several representative sections are presented (Fig. 49 ), as the relative proportions of the different lithofacies may vary stratigraphically. Thus the Kinlocheil Cross-Bedded Quartzite is characterised by a greater proportion of lithofacies 4 and 5 relative to 6 than, for instance, the Druim Fada Quartzite which is almost entirely composed of lithofacies 6, in association with thinly-bedded siltstones.

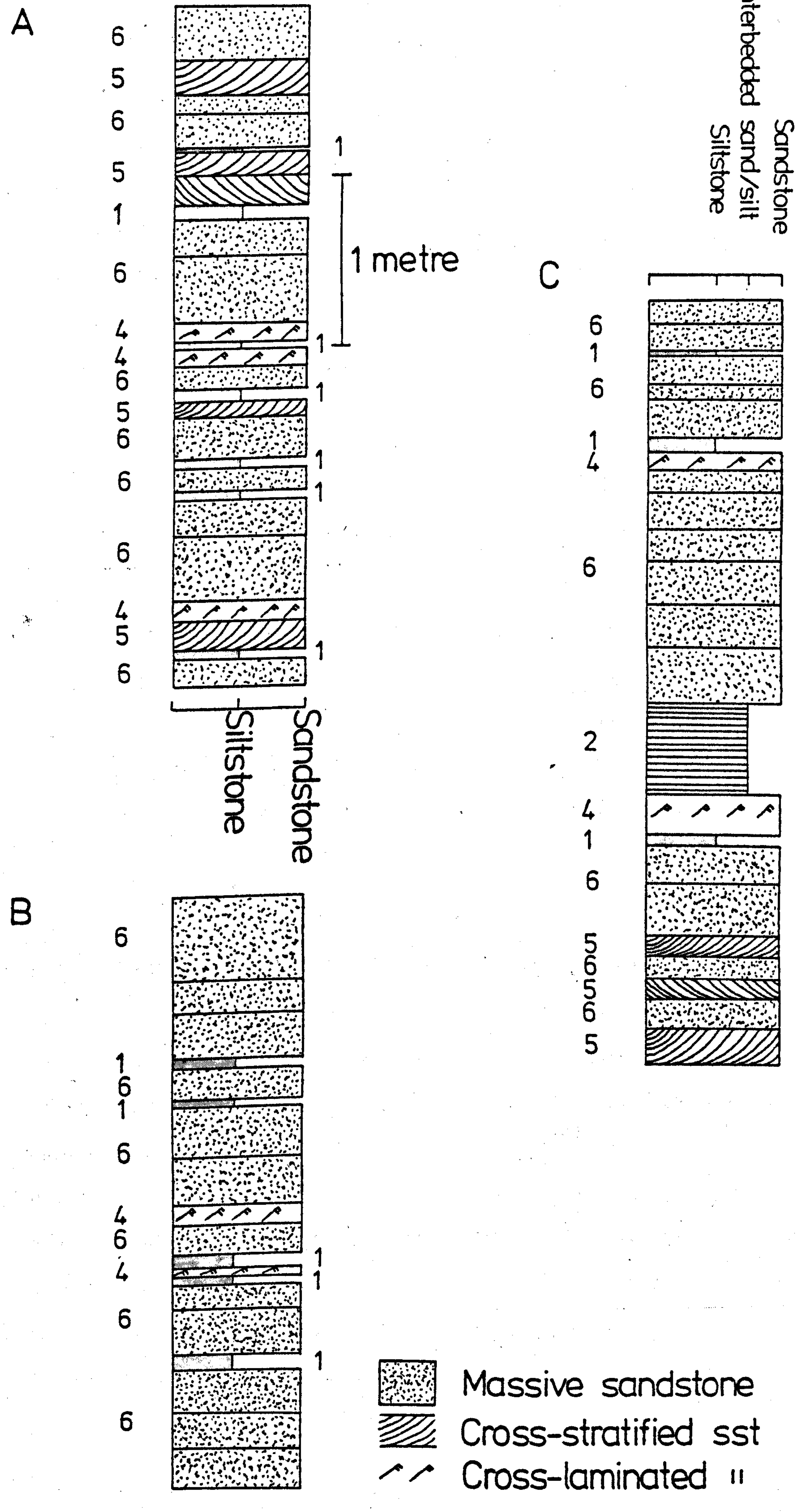
Figure 49: Representative logs of Association A:

A: Kinlocheil Cross-Bedded Quartzite (NM 961771).

B: Druim Fada Quartzite (NN 062821).

C: Glen Garvan Psammite (NM 984817).

Interpretation: Dominantly deposition of sand from the migration of laterally extensive sand waves. Occasional deposition of silt during periods of high suspended sediment concentration.


A
B
C
1 metre
Sandstone
Siltstone
nterbedded sand/silt
Massive sandstone
Cross-stratified sst
Cross-laminated "

### 7.3.4. Association D

The interbedded sandstones and siltstones of lithofacies 2 dominate this association (Fig. 50) and occasionally include massive sands of lithofacies 6, and horizons of siltstone which may be up to several metres thick. This association corresponds to:

Kinlocheil Banded Quartzite
Glen Suileag Banded Psammite
Stronchregan Mixed Assemblage
Inverscaddle Psammite

Examples of cross-lamination were never seen in either the Kinlocheil Banded Quartzite or the Stronchregan Mixed Assemblage, but were occasionally observed in both the Glen Suileag Banded Psammite and the Inverscaddle Psammite.

### 7.3.5. Association E

This is characterised by the predominance of cross-laminated and cross-stratified sandstones, in association with thinly-interbedded siltstones. Massive sandstones are occasionally present. The diagnostic feature of this association is the ubiquitous presence of "herring-bone" structure. Flat-topped rippled surfaces described earlier (p. 157 ) are also present within the association. Volumetrically the association is the least important of all five, and is only developed at two localities alongside the A830 road 1-2 kilometres E of Fassfern (NN 03557840 and NN 04807840). At both S0/S1 is sub-horizontal and despite the sparseness of intervening exposures the two are inferred to occur at broadly the same stratigraphic horizon.

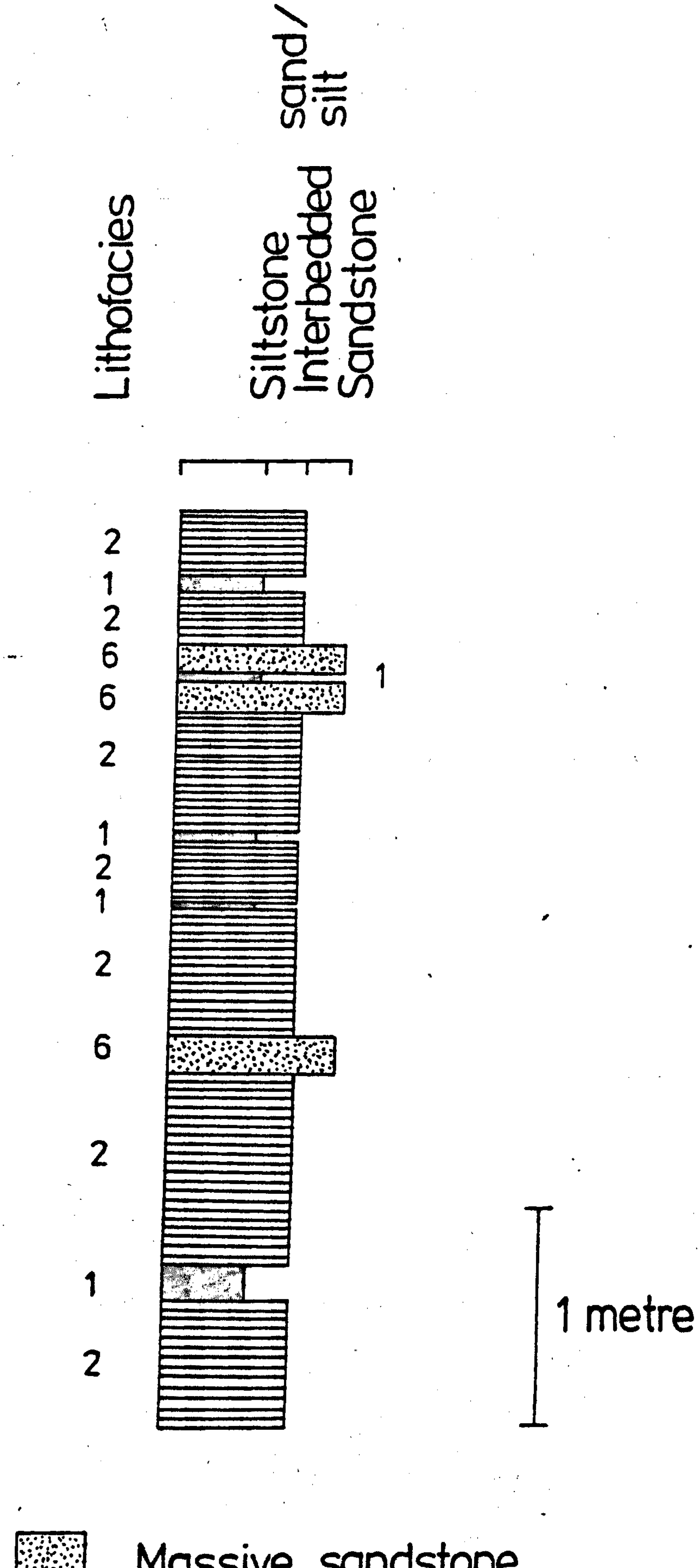



Figure 50: Representative log of Association D (Kinlocheil Banded Quartzite, NM 964804 ).
<u>Interpretation:</u> Dominantly rhythmic deposition of sands and silts on a centimetric scale. Occasional periods of high suspended sediment concentration give rise to thicker layers of silt, and rare massive sands are due to the migration of sand waves.

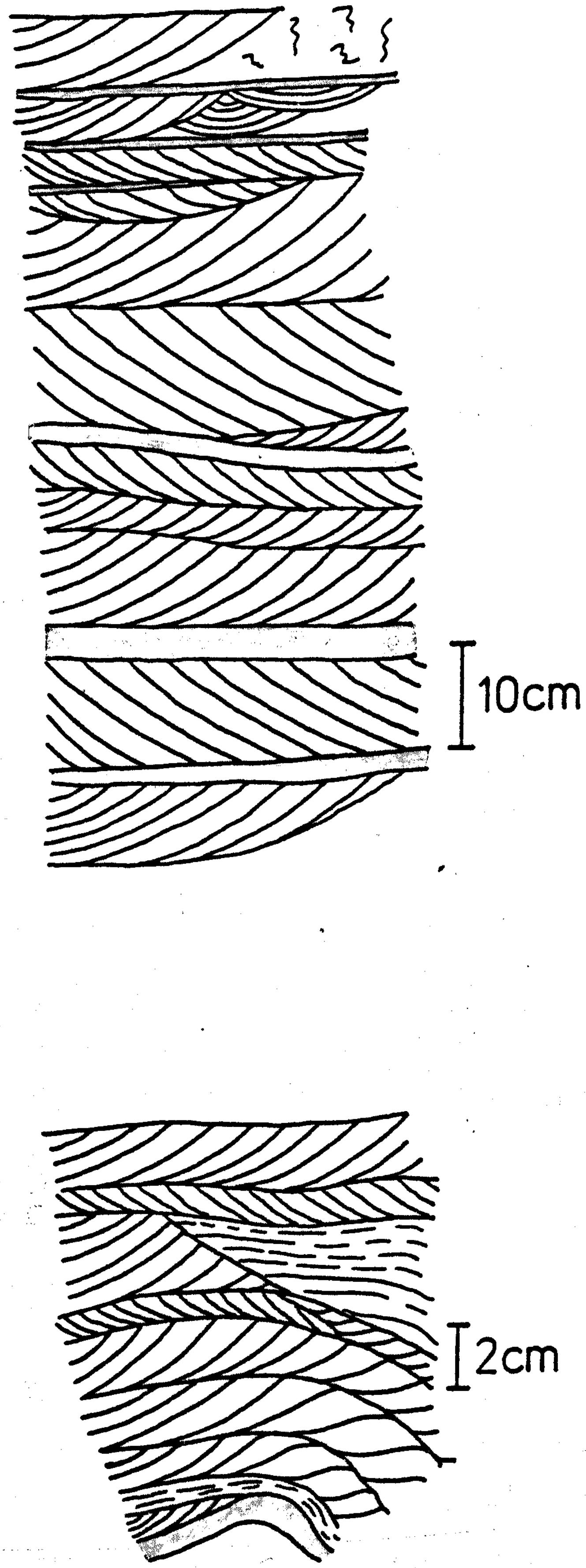



Figure 51: Diagnostic sedimentary structures of Association E (Cona Glen Psammite, NN 035784).

Sedimentary structures from these localities are depicted in Fig.51 and Plates 66 and 67 .

Fig.52 presents a generalised vertical section through the stratigraphic sequence of the Loch Eil Division, and shows the relative volumetric importance and order of occurrence of the five lithofacies associations.

## 7.4. ORIGIN AND FORMATION OF CALC-SILICATES

### 7.4.1. Occurrence

The field appearance and mineralogy of the calc-silicates have been described in detail in Chapter 3. Their mineralogy indicates that they represent original calcareous horizons within the sediments. They are common within the study area, and are present within most of the lithostratigraphic units described in Chapter 2. There would, however, seem to be a strict lithological control on their occurrence, as in the Loch Eil area, they are not present within siltstones nor within those highly siliceous sandstones which have been described as quartzites in their metamorphic state.

### 7.4.2. Morphology

The calc-silicates are morphologically highly variable. They tend to be best developed within well to thickly-bedded sands, ranging in size from elliptical pods (Plate 76 ) to more elongate bands which may be up to 15 centimetres thick and 1.5 metres long in the plane of S0/S1. More typically they form bands 20-25 centimetres long and 5-6 centimetres thick. Contacts with the enclosing sediments are usually sharp. They may be preferentially concentrated along certain

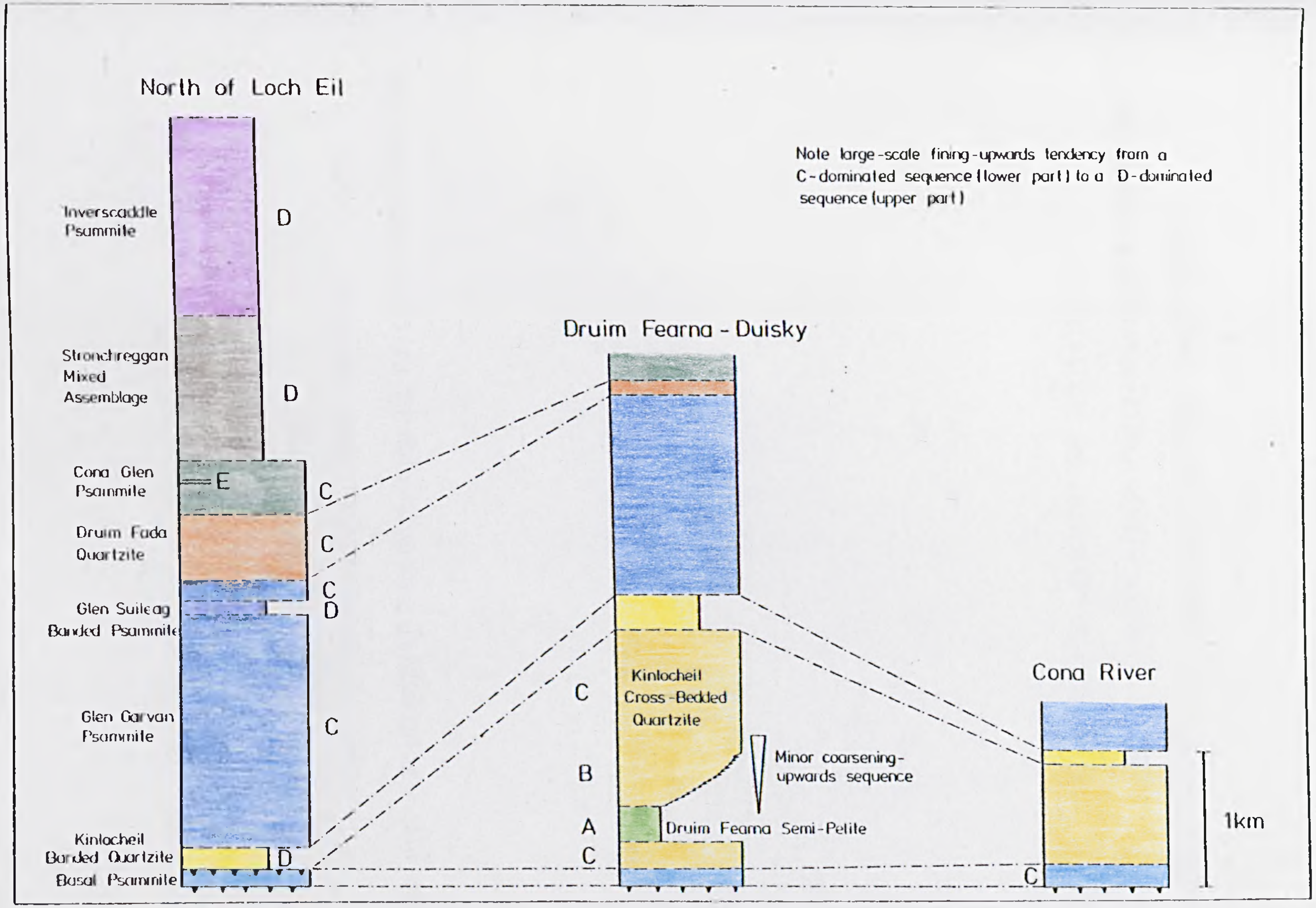



Figure 52: Generalised vertical section through the Loch Eil Division, showing the relative volumetric importance and order of occurrence of the five lithofacies associations.

Plate 76: Elliptical calc-silicate pod 'overgrowing' foreset laminae within cross-stratified sandstone. (Glen Garvan Psammite, NM 94957470).

Plate 77: Preferential concentration of calc-silicate pods and bands along particular sandy horizons. (Glen Garvan Psammite, NM 946744).

beds (Plate 77), occurring as mutually adjacent pods or bands which occasionally appear to be in the process of amalgamating to form one continuous calc-silicate horizon. Where lithologies rapidly alternate between sandstone and siltstone, as is characteristic of lithofacies 2, the calc-silicates form very thin laterally extensive horizons (Plate 78). Where the calc-silicates are comparatively poorly-developed they take the form of diffuse wisps or blebs which may only be 1-2 centimetres long.

### 7.4.3. Origin

It is essential to determine whether the calc-silicates are primary in origin, and originated as beds of calcareous material deposited at the same time as the enclosing sediments, or secondary features formed during diagenesis. The following observations are relevant with regard to this:

(a) The calc-silicates are nearly always seen to be lenticular in form; this is in marked contrast to 'normal' beds of sandstone and siltstone which are rarely observed to "wedge-out".

(b) At several localities (e.g. NM 949747) calc-silicate pods appear to 'overgrow' the foresets of cross-stratified sandstones (Plate 76). Foresets may also serve as the focus for the preferential growth of wisps of calc-silicate material (Plate 79). Small calc-silicate pods are also recorded from cross-laminated sandstones where they mimic the form of trough sets (Plate 80).

(c) The association of calc-silicate pods with undeformed sedimentary structures, and bedding which does not appear to have been particularly attenuated, suggests that the pods are not the

Plate 78: Thin laterally extensive calc-silicate bands with interbedded sandstone and siltstone of lithofacies 2. (Glen Suileag Banded Psammite, NN 03258340).

Plate 79: Preferential growth of wisps of calc-silicate material along foreset laminae. (Glen Garvan Psammite, NM 94957470).

Plate 80: Small calc-silicate pods mimic the form of trough sets within cross-laminated sandstone. (Glen Garvan Psammite, NM 946744).

Plate 81: Soft-sediment deformation: calc-silicate has 'loaded' downwards into underlying sandstone. (Glen Garvan Psammite, NM 988788).

result of the boudinage of originally more continuous layers, but are an original sedimentary feature.

All these observations are consistent with the interpretation that the calc-silicates are secondary in origin and represent a series of concretionary pods and layers formed during the diagenesis of the enclosing sands.

### 7.4.4. Mode of Formation

This is difficult to assess with any degree of certainty since much valuable textural information has undoubtedly been obscured by recrystallisation during metamorphism. However, there would appear to be two broad alternatives with regard to this:

(a) The calc-silicates are the result of the segregation of detrital carbonate already present within the sediments. As noted by Ramberg (1952, p.222), because of surface energy differences, free energy will be less if these constituents occur in clusters rather than finely divided and disseminated; hence aggregates will form in time. It seems likely that grains of detrital carbonate would be concentrated in sandy lithologies, and this would account for their absence from the siltstones of the study area.

(b) The calc-silicates formed as a result of the introduction of carbonate by circulating pore fluids, and consequent precipitation in the pore spaces of the host sediment (Lippman 1935; Krumbein and Garrels 1952; Hayes 1964; Deegan 1971). The greater pore space available within sandstones would similarly result in the preferential formation of calc-silicates in sandstones with respect to siltstones (Raiswell 1971).

The relative absence of calc-silicates from quartzite with respect to psammite is, however, difficult to explain if the carbonate was entirely derived from circulating pore fluids, since there is unlikely to be any significant difference in pore space between the two lithologies. The distribution of calc-silicates is easier to explain if they resulted from the segregation of detrital carbonate, since this might be expected to have been considerably less common within the quartzites due to their relative mineralogical maturity. This consideration suggests that the formation of the calc-silicates might therefore be attributed to the segregation of detrital carbonate. The relative abundance of calc-silicates both on outcrop scale, and within the lithostratigraphic succession, is thus likely to be a direct reflection of the amount and distribution of detrital carbonate originally present within the sediment.

### 7.4.5. Time of Formation

The age of formation of the calc-silicates must be estimated in relation to the time of deposition of the enclosing sediments. With regard to this, Pantin (1958) suggested three broad age divisions:

(a) syngenetic - formed at the time of deposition of the enclosing sediments.

(b) diagenetic - formed in the enclosing sediments while they are still soft and unconsolidated.

(c) epigenetic - formed after the consolidation of the enclosing sediments.

Calc-silicates which have been observed to 'overgrow' cross-stratification and cross-lamination are clearly not syngenetic in origin. Pantin (op cit) suggests that syngenetic concretions which grew at the sediment-water interface would show a different sense of symmetry to those which grew entirely within the sediment. The upper surface of such concretions would be growing under different conditions to those at the lower boundary, resulting in a probable asymmetric form. The fact that most calc-silicates, particularly the more ellipsoidal ones, show a plane of symmetry approximately parallel to bedding suggests again that few, if any, are syngenetic.

Several localities display laminations which diverge around ellipsoidal calc-silicates (Fig. 53 ). Unfortunately it is never possible to trace such laminations into and through calc-silicates, but the overall configuration bears a strong resemblance to that figures by Raiswell (1971, p.153, and Fig. 54 of this study). He suggests that the growth of the concretion would preserve by cementation the sedimentary laminae, which, within the first-formed central zones of the concretion, would be displayed as subparallel structures. As compaction progressed, laminae in the host sediment would be deformed around the relatively rigid cemented concretion to give the form in Fig. 53 . This therefore implies that the calc-silicates in question grew prior to compaction and consolidation of the sediments, and are therefore diagenetic in origin. At Kinlocheil (NM 98807875) calc-silicates appear to have been involved in soft-sediment deformation (Plate 81). This suggests that the sediment was still water-saturated even after the formation of the calc-silicates which are therefore likely to be diagenetic in origin. Given that conditions of permeability within massive sands were probably nearly isotropic, it might be expected that some of the calc-silicates would be virtually spheroidal in form. The fact

Figure 53: Laminations diverging around a calc-silicate pod (Glen Garvan Psammite, NM 94957470).

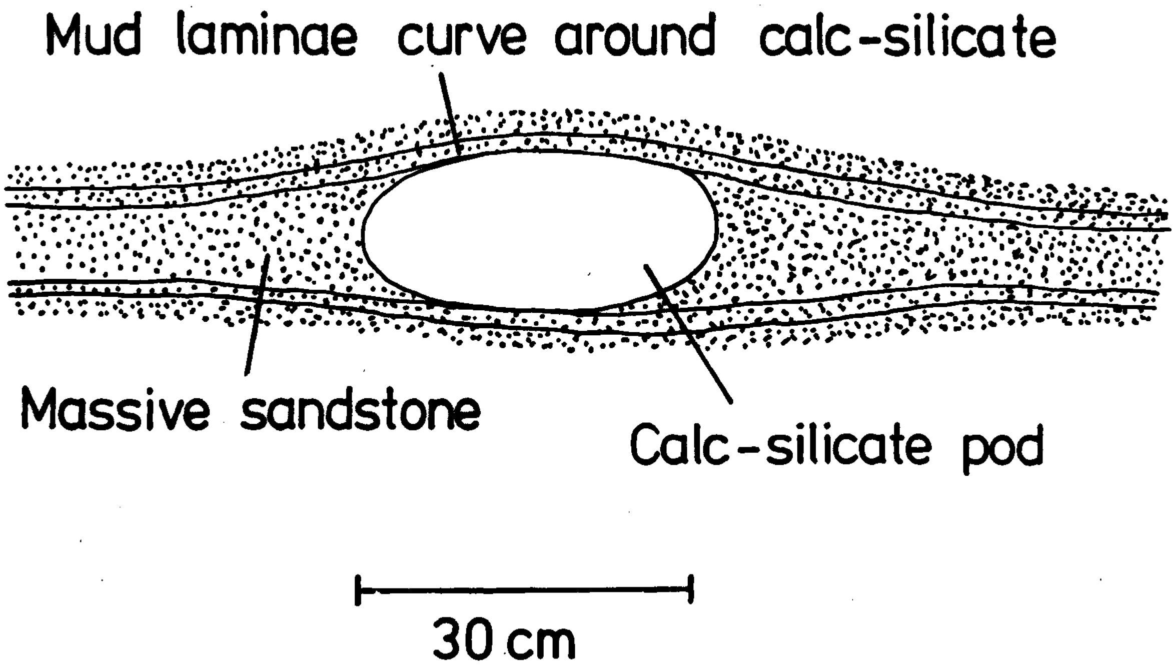
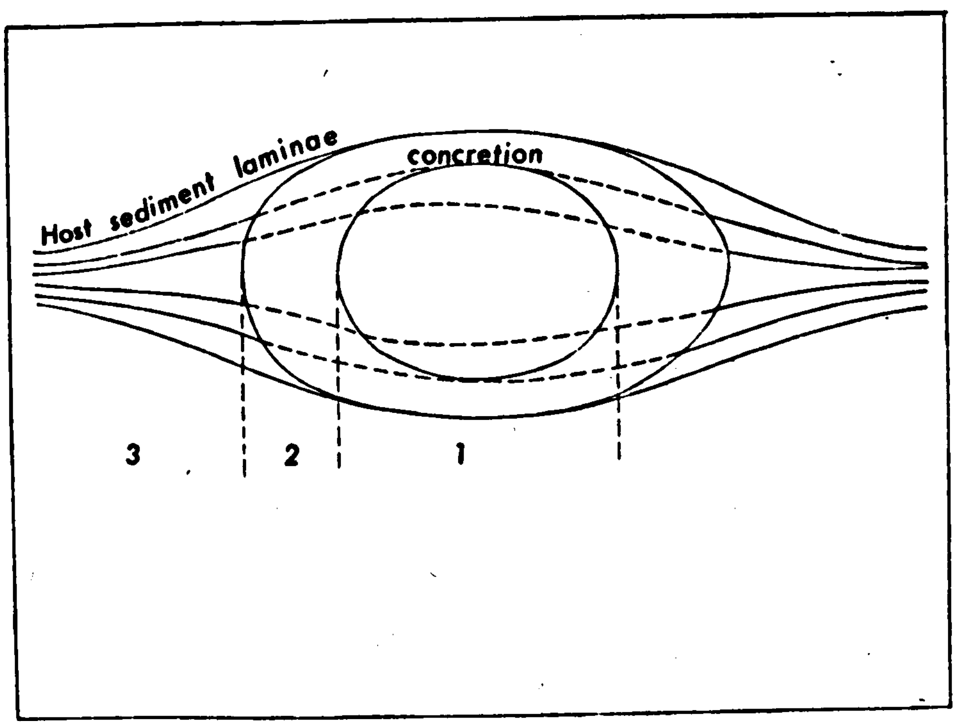



Figure 54: The growth of a deformed laminae structure. 1: initial cementation of uncompacted sediment; 2: sediment compacts around cemented concretion; 3: the growing concretion cements the surrounding compacted sediment, whilst the host sediment laminae are further compacted around the concretion. (Taken from Raiswell 1971).



that none are ever seen to take this form suggest some degree of compaction after their formation, again implying a diagenetic origin.

It is difficult to establish precise criteria for the recognition of epigenetic calc-silicates: those which form thin elongate bands, and those which are ellipsoidal but not associated with deformed laminae, may possible be epigenetic in origin.

In conclusion, where there is any firm indication of the time of formation of individual calc-silicates, a diagenetic origin is indicated. It seems likely that most of the calc-silicates were formed under these conditions, although a prolonged history of formation which overlaps all three of Pantin's categories cannot be ruled out.

## 7.5. PALAEOCURRENT ANALYSIS

Reconstruction of palaeocurrent directions in relatively undeformed sequences depends on the basic assumption that the current originally flowed in the direction of maximum foreset inclination of cross-stratified and cross-laminated beds. Various techniques have been developed for the reconstruction of palaeocurrent patterns in tectonically deformed rocks (Ramsay 1961). Several major problems are encountered in any attempt to reconstruct palaeocurrents in rocks which have undergone polyphase deformation:

(a) Practical difficulties relating to the accurate measurement of foresets on hillside exposures where often only two-dimensional exposures are available.

(b) Considerable distortion may occur within beds during deformation, resulting in steepening and shallowing of foreset inclinations.

(c) The accuracy of results is strongly dependant on the type of folding. It is possible to rotate sedimentary units back to the horizontal position comparatively easily when they have only been deformed by flexural-slip folds, and it is also possible to account for rotation due to the plunge of fold axes.

If, however, the rocks have been deformed by similar folding, the tectonic axes 'a', 'b' and 'c', together with the amount of compression, must all be calculated accurately. Considerable errors may result by unfolding using methods only applicable to flexural folds. Difficulties are further magnified when the fold axes are non-cylindrical.

The first two of the difficulties listed above are unlikely to greatly constrain palaeocurrent analysis in the Loch Eil Division since it has already been indicated that despite the restricted nature of outcrops, sections normal to the current flow are the most commonly observed, and thus the direction of current flow is along the strike of the tilted bed. Furthermore, internal strain within the beds does not seem to have been marked, as there is little evidence of either steepening or shallowing of foreset inclinations. However, given that the Loch Eil Division has undergone polyphase deformation, and that major folds are commonly non-cylindrical, it would be unrealistic to attempt the accurate reorientation of cross-bedding throughout most of the area.

Reorientation of cross-bedding S of Loch Eil within the Basal Psammite, Druim Fearna Semi-Pelite and the Kinlocheil Cross-Bedded Quartzite does, however, seem possible given their relatively simple tectonic history. Tilting of bedding was largely accomplished during

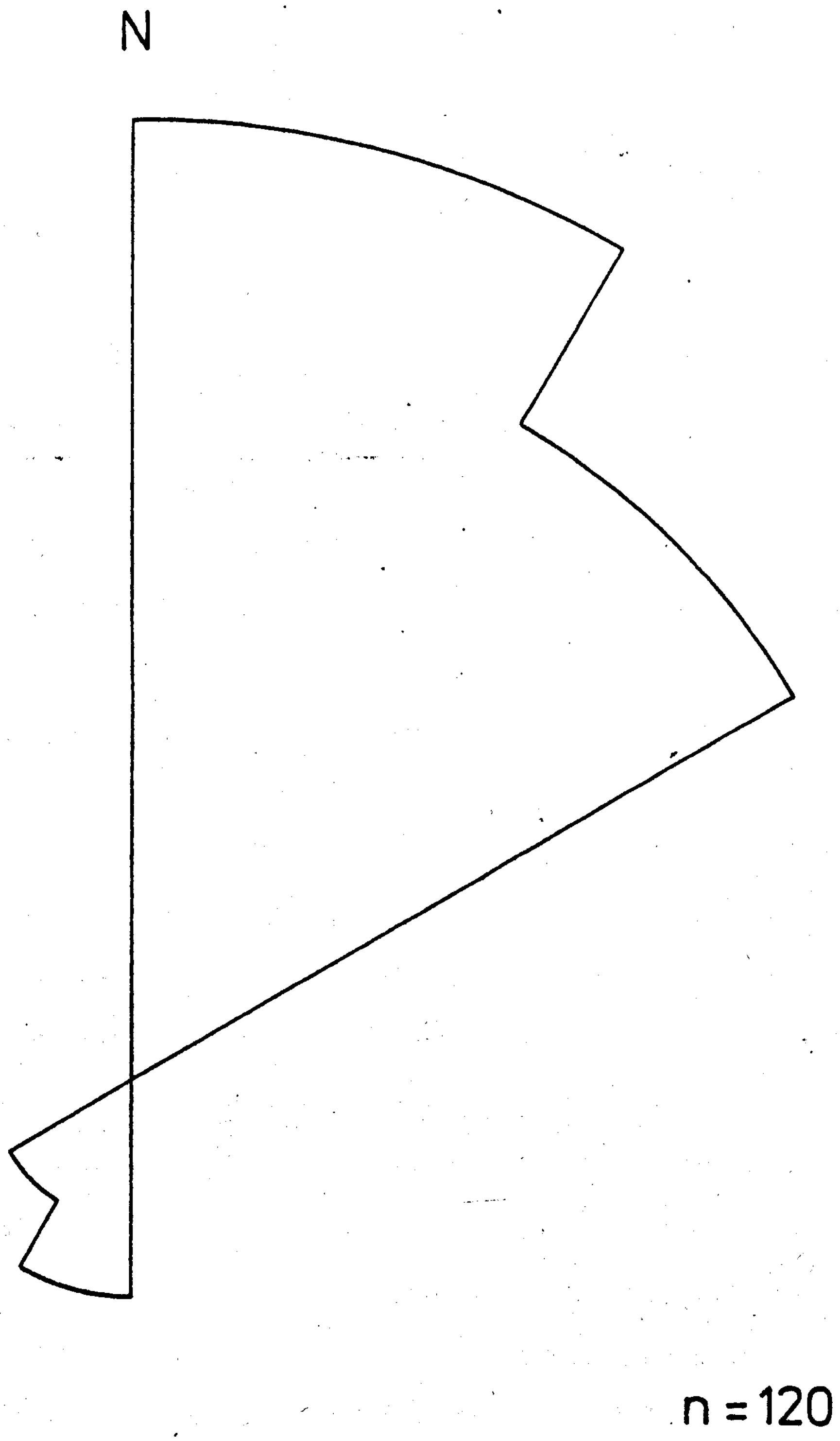



Figure 55: Rose diagram resulting from the reorientation of cross-bedding S of Loch Eil.

$D_1$ sliding (p. 73 ), and the effects of $D_2$ and $D_3$ major folding are mainly confined to the Glen Garvan Psammite E of the North Garvan River. It is considered that the reorientation of foresets by simple correction of tilt may closely approximate to the original direction of palaeocurrents. This process has been carried out and the resultant compass rose diagram is shown in Fig. 55 . It indicates that the majority of foresets were originally inclined to the NNE, and therefore that this was the dominant direction of sediment transport. A subsidiary mode to the SSW represents occasional reversals in the direction of sediment transport: the fact that these reversals arise from the presence of "herring-bone" bedding indicates that they may be attributed to tidal activity.

## 7.6. ENVIRONMENTAL INTERPRETATION

The most reliable method by which marine and non-marine sediments may be differentiated depends upon the identification of certain indicators which are controlled by the salinity and depth of seawater, such as marine body fossils, trace fossils, and certain geochemical parameters. Thus Reading (1978b, p.478) comments: ".... in the Precambrian, facies interpretation is much more difficult, even in clastic sediments. We are not always sure of distinguishing a shallow-marine environment from a fluvial one, let along lacustrine from marine ....". However, notwithstanding the absence of salinity indicators, a combination of other sedimentological data, such as sedimentary structures, lithofacies associations, palaeocurrent patterns, vertical sequences and sand body geometry and trend can provide a means of differentiating between the various depositional environments. Few of these parameters are by themselves diagnostic of any one particular environment, but grouped together they may be very effective in interpretation.

Any such interpretation must explain the following features of the sediments:

(a) The exclusively tabular bed geometry, and lateral persistence of individual sedimentation units.

(b) The widespread lateral and vertical uniformity, at least in two dimensions, of the lithofacies and lithofacies associations.

(c) The lack of any obvious cyclicity within the sediments.

(d) The total absence of any coarse clastic material such as pebble bands or conglomeratic horizons.

(e) The presence of thick sequences of quartzite which may be interpreted as mature polycyclic sands.

(f) The lack of any deep channelling; erosion surfaces, if present, must be inclined at such low angles to bedding as to be virtually undetectable by means of outcrop mapping.

(g) The occasional presence of "herring-bone" cross-bedding and wave-ripple cross-lamination.

It seems most likely that thick, dominantly sandstone, sequences such as that under discussion must have accumulated under either fluvial, deltaic or shallow-marine conditions, or a combination of these environments. Accordingly it is proposed to discuss briefly the major diagnostic features of each, and to compare them with those of the Loch Eil Division sediments.

### 7.6.1. Fluvial Systems

Reviews of sedimentation in fluvial systems are presented by Collinson (1978) and Walker (1979). Collinson (op cit, p.42) comments: "Alluvial sedimentary rocks are generally recognized by the absence of marine fossils, the presence of red coloration, unidirectional palaeocurrents and channels, and evidence for emergence such as the presence of palaeosols and mudcracks .... in the Pre-Cambrian where a marine fauna is lacking and soil processes were less active, it may be difficult to distinguish fluvial from shallow marine sediments". He subdivides ancient alluvial systems into 'ancient pebbly alluvium' and 'ancient sandy fluvial systems'. The sediments under discussion clearly bear no resemblance to the pebbly alluvium category, which is characterized by a substantial proportion of conglomerates, often developed in association with channelling and mud-flow deposits. In most sandy fluvial systems, two major facies associations may be distinguished: these are 'coarse' and 'fine' members which are usually interpreted as laterally accreted channel deposits and vertically accreted inter-channel deposits respectively. 'Coarse' member deposits are characterized by thin conglomeratic 'lag' deposits, trough cross-bedded sandstones together with cross-laminated and parallel-laminated sands. 'Epsilon' cross-bedding may be present, but is not particularly common. Silts and muds dominate the 'fine' member deposits and often include thin beds of sand which may be cross-bedded. The two members show a tendency to be organized into repeated fining-upwards cycles, due to the lateral migration of channels. The absence of such cycles, or any of their major components, from the Loch Eil Division sediments makes it unlikely that these rocks accumulated in a fluvial system.

## 7.6.2. Deltas

The major features of deltas are summarised by Elliott (1978) and Coleman and Wright (1975). Deltas are complex and morphologically highly variable, and consequently several models are available as an aid to the interpretation of ancient deltaic sequences. Thus Coleman and Wright (op cit) present a series of idealised vertical sections through deltas which all differ to some degree in the style and nature of deposition. Galloway (1975) erected a basic tripartite division of deltas, which recognized fluvial-, wave- and tide-dominated regimes. As deltas comprise an association of depositional environments, their recognition in ancient sequences requires the identification of a number of related facies associations. Elliott (op cit) identifies three major facies associations within the delta system: the delta plain and delta front facies associations, deposited whilst the delta is active, and the delta abandonment facies association. Delta plain sequences are generally very similar to those produced by the lateral migration of fluvial channels, and, as has already been stated above, it seems unlikely that they are responsible for the deposition of any great part of the Loch Eil Division succession. The delta front is represented by a relatively large-scale coarsening-upwards sequence which records the passage from fine-grained offshore or prodelta facies upwards into sandstone-dominated shoreline facies. This sequence results from progradation of the delta front, and successive abandonment of distributary channels leads to the repetition of such major coarsening-upward cycles, which are possibly the most diagnostic feature of deltaic deposition. Abandonment facies are thin laterally extensive horizons which are volumetrically of little importance.

Although all the individual lithofacies associations which have been described from the Loch Eil Division succession probably could be incorporated within a model of deltaic deposition, a comparison of their relative proportions and vertical arrangement (Fig. 52 ) with vertical sequences from both ancient and modern deltaic sequences (Elliott op cit) reveals certain major differences. Firstly, the deltaic sequences usually include a substantially greater proportion of silt and mud than is included within the vertical section for the study area. Secondly, and more importantly, there is an absence from the study area of the repeated coarsening-upwards sequences which are characteristic of deltaic deposits. The one major coarsening-upwards sequence present may superficially resemble such cycles, but its upper sandy member does not display any sedimentary features which might indicate the progradation of a complex of fluvial- or tidally-dominated distributary channels over delta-front sands and silts. The absence of such features, both from this particular sequence and elsewhere in the Loch Eil Division succession, indicates that these sediments are unlikely to be the direct result of deposition in a deltaic environment.

### 7.6.3. Shallow-marine environments

To date, the study of modern shallow-marine environments has failed to produce a generalised model for such deposits which might be applied to ancient sequences. This failure has resulted in part from the atypical nature of modern shelves compared with ancient shallow seas, since, following the Holocene transgression, most show some degree of disequilibrium. Thus while they provide valuable transgressive models, they are not directly applicable to ancient regressive or equilibrium situations. Despite the observation that modern shelf sands are generally not being derived from contemporary coasts, but are reworked "relict" deposits, numerous authors have attributed thick extensive sandstone formations to deposition in shallow-marine environments (Singh 1969; Anderton 1976; Levell 1980). These studies suggest that in the geological past great thicknesses of sand were deposited in actively aggrading shallow-marine environments.

Most of the lithofacies and lithofacies associations described earlier have ready analogues both in the modern shallow-marine environment (Belderson and Stride 1966; Kenyon 1970), and in ancient shallow-marine deposits (Johnson 1978 p.234). Furthermore, the characteristic features of the Loch Eil Division sediments (p.172) are all closely comparable with those of ancient shallow-marine successions, and there are a number of particularly striking points of comparison:

(a) The exclusively tabular bed geometry, and lateral persistence and uniformity of both small- and large-scale sedimentation units within the Loch Eil Division succession are common features of shallow-marine sediments deposited on stable, gently-sloping continental shelves (Levell op cit).

(b) The lack of any obvious cyclicity within the Loch Eil Division sediments is entirely consistent with deposition in a shallow-marine environment. In contrast to the relatively ordered cyclic nature of fluvial and deltaic sequences, shallow-marine deposits commonly lack any ordered vertical sequence of lithofacies associations (Anderton op cit, p. 435). This may be attributed to the more complex distribution and evolution of shallow-marine bedforms and sediments in response to a large number of interacting factors (Johnson 1978 p.208).

(c) Thick sequences of quartzite, similar to the Kinlocheil Cross-Bedded and Druim Fada Quartzites, are particularly common in ancient shallow-marine successions (Klein 1970; Swett 1971; Anderton 1976), and are attributed to the prolonged reworking of sands in near-shore zones. This process may also account, in part, for the absence of any coarse clastic material within the Loch Eil Division sediments.

(d) The apparent lack of deep channelling within the Loch Eil Division sediments is again consistent with deposition in a shallow-marine environment, since erosion surfaces, if present, are frequently broad flat features which only truncate bedding at low angles, and would consequently be difficult to detect by means of outcrop mapping.

(e) The local presence of "herring-bone" cross-bedding within the Loch Eil Division sediments indicates that at least part of the succession was deposited within a tidal regime, and this is clearly compatable with deposition in a shallow-marine environment.

### 7.6.4. Conclusions

The characteristic features of the Loch Eil Division succession are therefore considered to result largely from deposition in a shallow-marine environment. Clearly, however, all environmental interpretation must be generalized, and it is acknowledged that thin unrecognized fluvial or deltaic incursions may be present.

## 7.7. A DEPOSITIONAL MODEL

Given the recognition of a shallow-marine environment, it is clearly necessary to refine this interpretation into a more specific depositional model. Any such model must take account of the following features of the Loch Eil Division succession:

(a) The considerable thickness of sediment involved (c.6-7.5 kilometres), which implies that deposition occurred in an actively aggrading shallow-marine environment.

(b) The comparative uniformity of sediment type, which suggests that deposition occurred in a tectonically stable area, which was gently subsiding at a rate which kept pace with sedimentation.

(c) A large-scale fining-upwards tendency (Fig. 52).

Shallow-marine sedimentation may be subdivided into two distinct regimes (Swift 1976): a passive regime in which the shelf sand sheet is generated by erosional shoreface retreat (autochthonous sedimentation), and a more active regime in which river mouth bypassing causes deposition across the continental shelf (allochthonous sedimentation). The nature and characteristics of each of these regimes are outlined briefly:

Autochthonous regimes occur as a result of rapid transgression, during which river mouths generally cannot adjust to their combined river and tidal discharges as fast as required by the rise of sea level, and they become sediment sinks. In this case, there is little net input of sand from rivers onto the continental shelf, and shelf sands result almost entirely from the in situ reworking of fluvial and coastal plain deposits drowned by the transgression (Swift op cit, p. 314). The continental shelf around the British Isles is a good example of an autochthonous regime: the thin veneer of shelf sediment is essentially a result of the in situ reworking during the Holocene transgression of pre-Holocene sands and gravel deposits. Net deposition of sand on the shelf is negligible: considerable quantities of sand are trapped in river and estuary mouths, and indeed recent studies suggest that much coastal sand is derived from the shelf rather than the other way around (Kulm et al. 1975; Swift op cit; Wright 1977).

Allochthonous regimes occur as a result either of slow transgression or actual regression. During the former river mouths are more likely to equilibrate to their discharge, and bypass fine sand in quantities sufficient to result in net deposition on the shelf surface. During regressions, shorefaces become sediment sinks and actively advance seaward. Allochthonous shelves are generally situated adjacent to large rivers with high sediment loads, and are typically floored by fine sands, fine silty sands and silts. These sediments occur as seaward thinning sheets or strips of fine sand and silt oriented parallel to the coastline. Modern examples of allochthonous regimes are rare since all shelves show some degree of disequilibrium since the Holocene transgression. Nevertheless, the ten-fold reduction in the rate of eustatic sea-level rise experienced between 4000 and 7000 years ago (Milliman and Emery 1968 has resulted in a shift from autochthonous to allochthonous regimes in a number of shelf sectors (Curray 1964). Good modern examples of allochthonous regimes are the Georgia shelf, U.S.A. (Swift op cit), and the Niger shelf, W. Africa (Allen 1970).

It seems unlikely that the Loch Eil Division succession accumulated within an autochthonous regime. Modern-day examples only involve little, if any, net deposition on the continental shelf, and clearly other mechanisms are needed to explain the thickness of sediment involved. Several authors (Johnson 1978; Level 1980) have argued, however, that repeated transgressions should, theoretically at least, provide the most important mechanism for accumulating large quantities of sand on the shelf. They suggest that major sand bodies deposited during lower sea level stands (e.g. beach, barrier, fluvial and deltaic deposits) would be submerged during transgression and redistributed across the shelf by strong tidal currents and storm surge ebb currents. There is little evidence, however, that this has actually occurred within modern authochthonous regimes, even within the North Sea where vigorous tidal currents dominate sedimentation (Kenyon and Stride 1970). Furthermore, examination of the Loch Eil Division succession does not reveal any positive evidence of repeated transgressions, and neither does it seem likely that the level of tidal energy would have been sufficient to entirely rework relict gravels and redistribute large volumes of sand offshore. Accordingly, it is more probable that deposition of the Loch Eil Division occurred within an actively aggrading allochthonous regime, and this would more easily explain both the considerable thickness and relative uniformity of the succession.

In order to establish a model for the sedimentation of the Loch Eil Division it is therefore necessary to study the depositional patterns with known allochthonous regimes. Fig. 56 illustrates a simple model for the type of shallow-marine coastal-shelf environment within which the Loch Eil Division succession probably accumulated. This model is based on the work of Allen (1970) on the Niger Shelf, Howard and Reineck (1972) on the Georgia shelf, and Reineck and Singh (1973) on the Outer Jade and Nordergrunde areas of the North Sea. The salient features of this model are:

(a) The introduction of large quantities of sand and silt from a river system(s) onto the shelf, which is gently subsiding at a rate which broadly keeps pace with sedimentation.

(b) The dominance of cross-stratified, cross-laminated and evenly-laminated sands within the inner shelf, which evolve as a series of small laterally extensive sand waves during reworking and longshore transport by prevailing tidal currents. Small locales of silt deposition are also present within the coastal sands, from which they are separated by a transitional series of interbedded silts and cross-laminated sands. Such lateral heterogeneities in the style of deposition are common in shallow seas, since mud may accumulate in significant quantities in a variety of shallow-marine environemnts from the nearshore to deep offshore (McCave 1972), and thus laterally coexist with areas where active sand wave progradation is occurring. These locales of silt deposition receive influxes of sand as a result of periodic storm activity.

(c) The presence of a transitional zone between the coastal sands of the inner shelf and the silts and muds of the outer shelf.

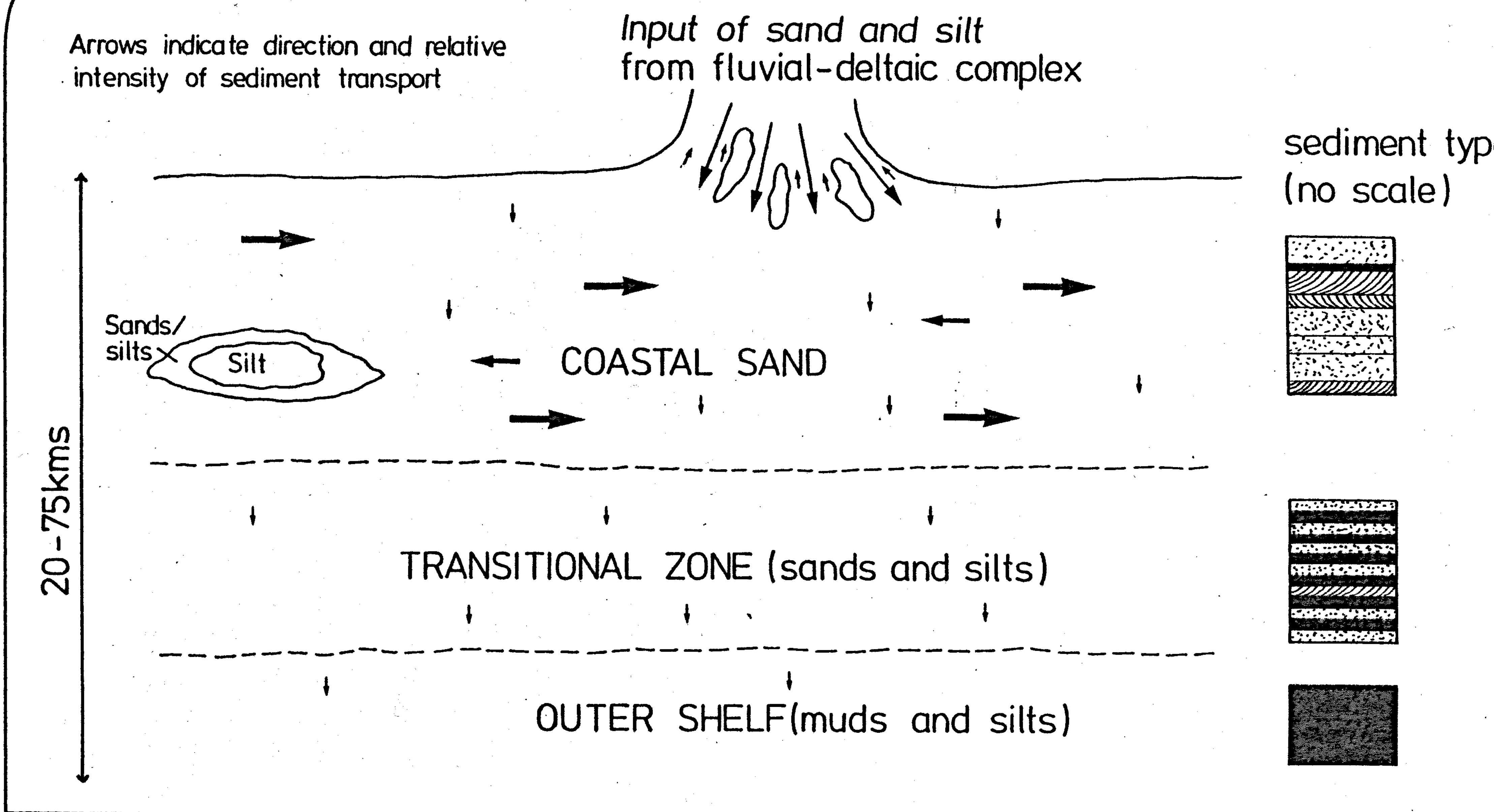



Figure 56: A depositional model for the sedimentation of the Loch Eil Division.

The transition zone is dominated by the interbedding of sand and silt on a scale of 1-3 centimetres. Silt is carried in suspension clouds through the high energy inner shelf area, and is deposited under relatively quiescent fair weather conditions. The source of sand is assigned to the coastal sand: during heavy storms and storm tides sediment is eroded in the nearshore zone and transported to the transition zone by retreating waves and ebb currents. Deposition of sand is largely by suspension, and less commonly by the migration of small-scale sand waves and ripples.

(d) An outer shelf zone where mud and silt are the dominant sediment type.

The volumetrically most important lithofacies associations within the Loch Eil Division succession are C and D (Fig. 52 ), and it is suggested that these correspond to the inner shelf coastal sand and transition zone respectively. Lithofacies A and B may be easily accommodated within the inner shelf environment as zones characterised by starvation of sand and relatively high rates of silt deposition. Lithofacies association E may also be incorporated within the coastal sand environment. The sedimentary structures recorded from this association are common to many tidal environments (De Raaf and Boersma 1971), and may represent the effects of a temporary shallowing in the shallow-marine environment, and the encroachment of a tidal flat or estuarine complex across the coastal sands. The outer shelf silt and mud does not appear to be represented in the Loch Eil area. Fig. 57 summarises the correlations which have been made between the lithofacies associations (and, by inference, the lithostratigraphy) and the shallow-marine

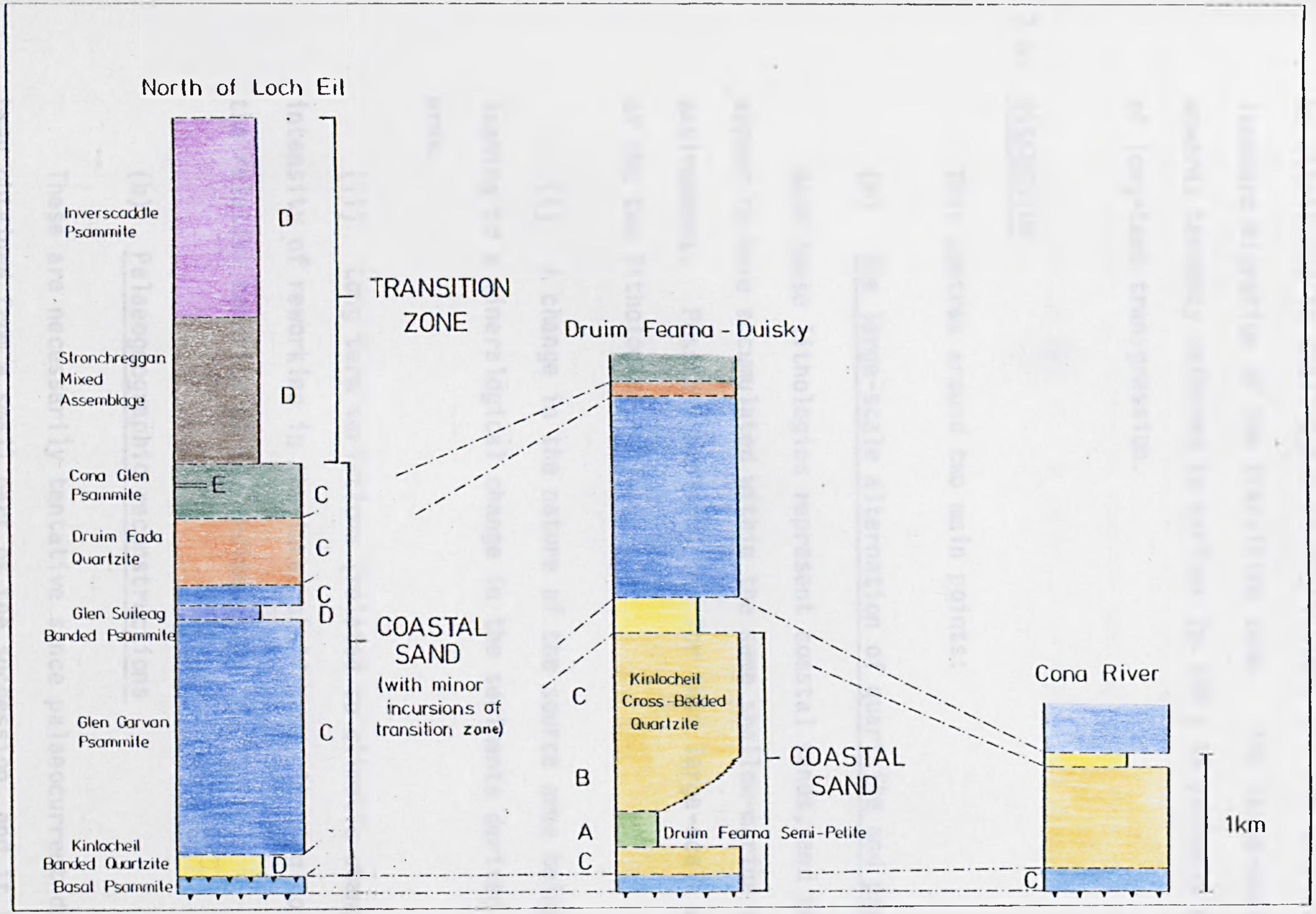



Figure 57: Correlations between the lithofacies associations/lithostratigraphy and the shallow marine sub-environments of Fig. 56.

sub-environments depicted in Fig. 56 . The alternation of associations C and D within the lower part of the succession may be attributed to small-scale transgressions and the temporary landward migration of the transition zone. The large-scale fining-upwards tendency referred to earlier (p. 178 ) is probably a reflection of long-term transgression.

## 7.8. DISCUSSION

This centres around two main points:

(a) The large-scale alternation of quartzite and psammite

Both these lithologies represent coastal sands, and therefore appear to have accumulated within the same shallow-marine sub-environment. Possible explanations for this large-scale alternation of the two lithologies include:

(i) A change in the nature of the source area being eroded, leading to a mineralogical change in the sediments derived from this area.

(ii) Long term variations (related to climatic change?) in the intensity of reworking in the inner shelf zone, leading to changes in the relative maturity of the sediments.

(b) Palaeogeographic reconstructions

These are necessarily tentative since palaeocurrent data has only been obtained from a small part of the succession, and it is uncertain whether this data is representative of the sequence as a whole. The available data suggests that the dominant direction of sediment transport within the lower part of the succession was to the NNE.

Within nearly all contemporary shallow-shelf environments the main direction of sediment transport is parallel to coastlines, and it therefore seems likely that this north-north-easterly transport direction reflects the activity of longshore currents flowing approximately parallel to a NNE-SSW trending coastline. It is not possible, however, to determine whether the landmass from which the sediment was derived lay to the NW or SE of the study area.

# CHAPTER 8: CONCLUSIONS AND DISCUSSION

## 8.1. CONCLUSIONS

These may be summarised thus:

(a) The metasedimentary rocks of the Loch Eil Division Moine of the Loch Eil area are composed of a series of psammites, quartzites and 'banded' units with minor semi-pelites and locally abundant calc-silicate pods and bands. Sedimentary structures are commonly preserved. The sequence has been subdivided on a lithological basis into ten lithostratigraphic units which total approximately 6,000-7,500 metres in thickness.

(b) Analysis of the lithofacies associations suggests that these metasediments accumulated in a shallow-marine environment within which tidal processes were operative. The formation of calc-silicates is attributed to the diagenetic segregation of detrital carbonate.

(c) The Loch Eil Division has been subjected to five phases of deformation (D1-D5):

(i) The first episode of deformation (D1) is represented by a series of tight to isoclinal folds with a penetrative axial planar foliation (S1) which is generally parallel to bedding (S0). D1 folds were probably originally recumbent in attitude and display a westerly sense of vergence. Westerly-directed D1 tectonic sliding occurred both at the base, and within the lower stratigraphic levels, of the Loch Eil Division and resulted in the interleaving of Glenfinnan and Loch Eil Division rocks in the Na h-Uamachan-Gulvain area.

(ii) Major folding occurred during D2, D3 and D4, and it seems likely that the interference of the D2 and D3 fold systems dominates the regional structure. Accompanying S2 and S3 fabrics are non-penetrative and restricted to semi-pelitic horizons.

(iii) The final episode of deformation to affect the area (D5) is represented by a series of brittle faults and thrusts.

(iv) Strain analysis based wholly on fold style suggests that the level of strain was at its peak during D1 and progressively declined thereafter. Of the three major events which are traceable throughout the area (D1, D2 and D3), D2 was characterised by a markedly heterogeneous strain pattern, whereas the level of strain during D1 and D3 remained relatively uniform across the area.

(d) Amphibolite facies metamorphism (M1) accompanied D1. Metamorphic grade was of upper to mid-amphibolite facies in the W of the area, and declined eastwards to mid- to low amphibolite facies. The dominant mineral assemblages and metamorphic fabric of the Loch Eil Division were acquired during M1. Subsequent metamorphic events (M2-M4) involved only the recrystallisation and local retrogression of M1 assemblages during metamorphism of amphibolite and greenschist facies.

(e) The Loch Eil Division is in tectonic contact with the underlying Glenfinnan Division Moine which, in the Loch Eil area, is composed mainly of semi-pelitic gneisses. These are commonly migmatitic and probably acquired their dominant mineral assemblages and metamorphic fabric during metamorphism of upper amphibolite facies. Petrographic analysis has not revealed any obvious compositional differences between the rocks of the two Divisions, although there are certain textural contrasts since the Loch Eil Division rocks are only locally gneissose and nowhere migmatitic, even though the two Divisions would appear to be of comparable metamorphic grade in the W of the area.

Structural observations along the eastern margin of the Glenfinnan Division suggest that it has been affected by all the deformational events which have been recorded within the Loch Eil Division. In addition, the Glenfinnan Division possesses indications of an earlier deformational history than that which is apparent in the Loch Eil Division: it seems likely that this earlier history involved the formation of at least part of the main foliation within the Glenfinnan Division, together with two phases of isoclinal folding.

(f) The granitic gneisses which crop out along the western margin of the Loch Eil area are considered to represent a series of deformed intrusive sheets, emplacement of which was pre- to syn-tectonic with respect to the first phase of deformation to affect the adjacent gneisses of the Glenfinnan Division.

(g) Two amphibolite suites are present within the Loch Eil area, both of which are considered to be igneous in origin. The 'early' amphibolites are widely distributed and are typically concordant with the dominant foliation within the enclosing metasediments. Emplacement of this suite is considered to be preto syn-tectonic with respect to D1. The 'late' amphibolites are restricted in their occurrence to a locality in the NW of the area: they are markedly discordant both to the foliation within the enclosing gneisses and adjacent early amphibolites. The intrusion of the late amphibolite suite appears to have occurred in the interval between D1 and D2 folding.

## 8.2. DISCUSSION

Discussion of the geological history of the Loch Eil area centres on two main topics:

(i) The status of the Loch Eil Division within the Moine Succession.

(ii) The nature and significance of the two major tectonic lineaments bordering the Loch Eil area, the 'Quoich line' and the Great Glen Fault.

Before embarking upon discussion of these topics it is pertinent to review the following geochronological constraints upon the timing of tectono-metamorphic events within the Loch Eil area:

(i) A 1050 $\pm$ 46 m.y. Rb-Sr isochron obtained from the Ardgour granitic gneiss at Glenfinnan (Brook et al. 1976). This was interpreted by them as dating the age of its formation as a migmatite (Dalziel 1963, 1966) during regional 'F2' folding (='D2' of this study). If the conclusions of Dalziel concerning the origin and timing of formation of the granitic gneiss relative to the regional structural sequence were correct, then it appeared that the main early metamorphism of the Moine, together with two phases of folding, occurred at about 1050 m.y., broadly coeval with 'Grenvillian' events in Norway and Canada.

Thus study suggests, however, that the granitic gneisses originated as intrusive bodies, emplacement of which was pre- to syn-tectonic with respect to the generation of the earliest structures within the enclosing gneisses of the Glenfinnan Division. There is no indication that the emplacement of these bodies was temporally associated with the D2 deformation, and the isochron obtained by Brook _et al_ (_op cit_) is here interpreted to date the formation of the gneissose fabric within the granitic gneiss and thus, by inference, widespread deformation and accompanying upper amphibolite facies metamorphism and migmatisation within the enclosing gneisses.

(ii) A 456 $\pm$ 5 m.y. U-Pb age obtained from igneous zircons within the Glen Dessary syenite (Van Breemen _et al_. 1979). This age was interpreted by them as dating the time of crystallisation of this igneous body, the emplacement of which appears to be pre- to syn-tectonic with respect to the regional 'F2' folds recognized by Johnstone _et al_. (1969) and Brown _et al_. (1970), and hence the 'D2' folds of this study. This age therefore provides a clear indication that the D2 deformation recognized within the Loch Eil area occurred during Caledonian orogenesis.

(iii) U-Pb, zircon and monazite ages of 445 $\pm$ 10 m.y. obtained from pegmatites in the Glenfinnan area (Van Breemen _et al_. 1974). The pegmatites are axial planar to, and thus syn-tectonic with, the 'F3' folds recognized by Johnstone et al. (op cit) and Brown et al. (op cit), and hence the 'D3' folds of this study. These ages provide further indications of Caledonian deformation within the Loch Eil area.

These ages are clearly suggestive of a polyorogenic history for the Glenfinnan Division rocks of the Loch Eil area; it is of central importance to establish whether the Loch Eil Division has undergone a similar sequence of tectono-metamorphic events.

### 8.2.1. The status of the Loch Eil Division within the Moine Succession

The geochronological evidence discussed above, when taken together with additional isotopic work on pegmatites and pelitic metasediments within both the Morar and Glenfinnan Divisions (Van Breemen et al. 1974, 1978; Brook et al. 1976; Brewer et al. 1979) clearly indicates that both Divisions underwent a complex poly-orogenic history. This may be interpreted in terms of initial deformation and metamorphism at ca. 1000 m.y., subsequent slow cooling and pegmatite emplacement at ca. 780-730 m.y., and large-scale reworking during the Caledonian orogeny (Powell et al. 1981). A large part of the Moine Succession may therefore be interpreted as a part of the Grenville belt of Norway and Canada. Direct indications of Precambrian tectonothermal events within the Loch Eil Division have not so far been forthcoming, and rely on the supposed presence of granitic gneiss bodies of the same age as the Ardgour granitic gneiss within the Loch Eil Division in Glen Garry, and the apparent structural unity between the Glenfinnan and Loch Eil Divisions suggested by Brown et al. (op cit.). Debate in recent years has focussed on the status of the Loch Eil Division within the Moine Succession, in particular whether it shares a common polyorogenic history with the Morar and Glenfinnan Divisions (Harris et al. 1978; Fettes 1979), or whether it represents a post-Grenville 'cover' sequence which has only undergone Caledonian orogenesis (Lambert et al. 1979; Piasecki 1980; Piasecki et al. 1981; Winchester et al. 1981).

Little detailed evidence has been published in support of the latter hypothesis and the present study is the first to conduct a detailed examination of the relationship between the Glenfinnan and Loch Eil Divisions in their type areas. The contact between the two Divisions in the Loch Eil area is sharp and planar and appears to be tectonic: there is little indication, however, of any substantial displacement across this contact, and it is therefore unlikely that the two Divisions are separated by a tectonic break of the same magnitude as the Sgurr Beag Slide which separates the Morar and Glenfinnan Divisions (Rathbone 1980; Powell et al. op cit). There is no direct evidence that the two Divisions were originally separated by an unconformity, which, if it existed, must have been so tectonically modified as to be unrecognizable, and similarly no suggestion that the two Divisions are linked by a sedimentary passage, contrary to the opinion of Fettes (op cit). Attempts to establish the original nature of the relationship between the two Divisions must therefore rest on the correlation of tectonothermal events across their contact, and the isotopic dating of the M1 metamorphism within the Loch Eil Division.

Evidence has been presented which suggests that the members of the Glenfinnan Division in the Loch Eil area were already either partially or wholly deformed prior to the main phase of D1 recumbent folding and tectonic sliding within the Loch Eil Division, and therefore that the structural relationships between the two Divisions are not as simple as considered by previous workers. The observed superimposition of structure might result from successive folding during a single progressive deformation, with the onset of deformation occurring rather earlier in the relatively ductile migmatitic rocks of the Glenfinnan Division than in the more competent non-migmatitic Loch Eil Division rocks. In this context the terms 'infrastructure' and suprastructure' might be loosely applied to the two Divisions. An alternative explanation is that the superimposition of structure reflects a 'basement-cover' type relationship between the two Divisions, and that Phase 1 and 2 structures represent the effects of Precambrian orogenesis in the Glenfinnan Division. Two alternative models for the evolution of the geology of the Loch Eil area are therefore presented, in Tables 12 and 13. The model outlined in Table 12 envisages that the two Divisions share a broadly common tectonometamorphic history, whereas that presented in Table 13 suggests that their relationship is one of basement and cover.

The'basement-cover' model is considered to be the more attractive, for several reasons. The westerly vergence displayed by D1 folds, and the suggested westerly sense of displacement along D1 tectonic slides in the Loch Eil area, are both consistent with early Caledonian deformation elsewhere in the N Highlands (Mendum 1979; Rathbone 1980), but less easily reconciled with the apparent easterly vergence of Precambrian fold nappes within the Morar Division (Ramsay 1963; Powell 1974). Furthermore, the

| Orogeny and timing of tectono-metamorphic events | | Geological events within the Glenfinnan - Loch Eil area | |
|---|---|---|---|
| | | D5 / M4 | |
| | | D4 } M3 | Reworking of rocks which were probably flat-lying after Precambrian deformation into variable attitudes. Accompanying metamorphism within the Loch Eil Division only resulted in the recrystallisation and local retrogression of M1 Precambrian mineral assemblages. |
| <u>CALEDONIAN</u> | c. 445 | D3 } | |
| | c. 456 | D2 / M2 | |
| | | Intrusion of late amphibolites (post 560m.y. according to Smith 1979). | |
| | 730-780 | Slow cooling and emplacement of 'Morarian' pegmatites within the Glenfinnan Division. | |
| <u>GRENVILLIAN</u> | 1000-1050 | D1 / M1 | Structurally polyphase event, involving the syn-tectonic migmatisation and early formation of two phases of recumbent folds within the relatively ductile rocks of the Glenfinnan Division, and subsequent recumbent folding and tectonic sliding within the more competent non-migmatitic Loch Eil Division rocks. Dominant mineral assemblages formed during metamorphism of upper amphibolite facies.<br>Pre- to syn-tectonic emplacement of granitic gneisses and early amphibolites. |
| | | Deposition of Glenfinnan and Loch Eil Divisions as a largely continuous and conformable sedimentary sequence. | |

Table 12: A common history for the Glenfinnan and Loch Eil Divisions

| Orogeny and timing of tectono-metamorphic events | | Geological events within the Glenfinnan - Loch Eil area | | |
|---|---|---|---|---|
| | | D5 / M4 | | Deformation and metamorphism of the Loch Eil Division entirely Caledonian in age |
| | | D4 } M3 | | |
| | c.445 | D3 } M3 | | |
| CALEDONIAN | c.456 | D2 / M2 | | |
| | | | Emplacement of late amphibolites | |
| | | DI/MI | Recumbent folding and tectonic sliding, resulting in the interleaving of 'basement' and 'cover' rocks. Initial metamorphism of the Loch Eil Division under amphibolite facies conditions. | |
| | | | Pre- to syn-tectonic emplacement of early amphibolites. | |
| | | Deposition of the Loch Eil Division unconformably upon the Glenfinnan Division. | | |
| | 730-780 | Slow cooling and emplacement of 'Morarian' pegmatites within the Glenfinnan Division. | | |
| GRENVILLIAN | 1000-1050 | Major deformational event, involving two phases of recumbent folding, accompanied by syn-tectonic migmatisation and upper amphibolite facies metamorphism which produced the dominant gneissose fabric of the Glenfinnan Division. | | |
| | | Pre- to syn-tectonic emplacement of granitic gneisses. | | |
| | | Deposition of the Glenfinnan Division | | |

Table 13: A 'basement-cover' model for the Glenfinnan and Loch Eil Divisions

basement-cover model more easily accounts for the general restriction of migmatitic and gneissose fabrics to the rocks of the Glenfinnan Division, and the apparent lack of any transitional zone between gneissose and non-gneissose lithologies, even though the two Divisions appear to be of comparable metamorphic grade in the W of the Loch Eil area. However, this model rests heavily on the assumption that the 'Loch Eil Division' rocks which enclose the granitic gneisses in Glen Garry are unlikely to be stratigraphically equivalent to the Loch Eil Division in its type area. Alternatively these granitic gneisses may have migrated to their present position subsequent to emplacement at a lower tectonic level (Winchester 1974; Piasecki et al. 1981). The general conclusion of this study with respect to the nature of the relationship between the two Divisions is therefore that although the available evidence might favour the basement-cover model, further mapping is clearly needed to clarify this issue.

If the basement-cover model is correct, then strong analogies may be made with the geology of the Grampian Highlands E of the Great Glen, where Piasecki (1980) and Piasecki et al. (op cit) record a complexly deformed gneissose 'basement' (the Central Highland Division) which appears to be overlain unconformably by a less deformed and less highly metamorphosed 'cover' sequence (the Grampian Division). These authors suggest the following tectono-stratigraphic correlations:

| | NW HIGHLANDS | GREAT GLEN FAULT | GRAMPIAN HIGHLANDS |
|---|---|---|---|
| POST-GRENVILLE 'COVER' | Loch Eil Division | | Grampian Division |
| Inferred unconformity | | | |
| GRENVILLE 'BASEMENT' | Glenfinnan Division | | Central Highland Division |

The lithological similarities between the Grampian Division and large parts of the Loch Eil Division have already been noted (Piasecki _op cit_; Piasecki _et al. op cit_), and comparison of the structural and metamorphic history of the Loch Eil Division of the Loch Eil area with that recently proposed for the Grampian Division of the Corrieyaireck-Killin area (Whittles 1980; Haselock 1982) reveals considerable similarities with respect to the style and orientations of successive deformations, and the timing and grade of associated metamorphic events. Comparisons such as these are dependant, however, on the assumption that the history of the Great Glen Fault is restricted to a relatively simple sinistral displacement of c. 160 km (Piasecki _et al op cit_. Fig. 1) whereas it may be a structure of more fundamental significance across which few correlations may be made.

### 8.2.2. Nature and significance of major tectonic lineaments

#### 8.2.2.1. The 'Quoich line'

The term 'Quoich line' is used here with reference to a narrow zone broadly coincident with the junction between the

Glenfinnan and Loch Eil Divisions, across which there are marked changes in lithology and attitude of foliation, and along which there is a notable concentration of a variety of meta-igneous and igneous rock types.

Brown et al. (1970) considered that the large-scale changes in the attitude of foliation across the Quoich line (i.e. the 'steep' and 'flat' belts of Leedal) are related to variations in the intensity of regional F2 folding. Thus F2 folds within the Glenfinnan Division are typically steeply-plunging isoclinal structures, whereas those in the Loch Eil Division are commonly gently-plunging tight to open folds. This variation in the style and attitude of F2 folds is attributed by Brown et al. (op cit, p. 323) to "non-uniform flattening of layers" during F2 folding, as a result of the obvious lithological differences between the two Divisions. It is clear from a study of the structural data presented in Fig. 7 and Enclosure 4 of this study that there is no obvious systematic variation in either the attitude of foliation, or the style and attitudes of D2 folds, within the Loch Eil Division with respect to the boundary with the Glenfinnan Division. The formation of the 'steep' belt would therefore appear to have been accomplished largely within the Glenfinnan Division. Nevertheless, the only areas where the Glenfinnan Division rocks of the Loch Eil area are steeply-foliated (Cona Glen and Nah-Uamachan) are both areas of D2 folding and this tends to confirm the genetic relationship between D2 folding and the formation of the 'steep' and 'flat' belts. The observation that the degree of heterogeneity of the D2 deformation increases westwards within the Loch Eil Division is also consistent with the interpretation of Brown et al. (op cit). Since the D2 folding occurred at c. 456 m.y. (Van Breemen et al. 1979) it is clear that the 'steep belt' is a Caledonian feature (Johnson et al. 1979, Fig. 3).

Of more fundamental significance is the concentration along the Quoich line of a variety of meta-igneous rock-types:

(i) The granitic gneiss belt closely follows the Glenfinnan-Loch Eil Division boundary between Strontian and Glen Moriston, only deviating from the Quoich line in the Glen Garry area.

(ii) The early amphibolite suite is generally restricted to the Loch Eil Division (Smith 1979).

(iii) The late amphibolite-metagabbro suite is concentrated along the Quoich line (Smith op cit. Fig. 5).

(iv) The late to post-tectonic felsic porphyrite and micro-diorite suites are similarly concentrated along the Quoich line (Smith op cit, Figs. 2 and 3).

This is suggestive of a major lineament which controlled the emplacement of intrusive material over a long period of time during the history of the N. Highlands (Smith op cit, p. 696). The coincidence of this lineament with a major lithological boundary might suggest a tectonic control on sedimentation. Current models for the sedimentation and tectonic evoluation of the Scottish Highlands stress the importance of fault-controlled deposition (Rathbone 1980; Anderton 1982; Stewart 1982; Soper and Barber 1982) and in the context of these models it seems possible that the Quoich line is represented at depth by a structure which originated, prior to the Grenville event, as an extensional fault within Lewisian basement. Subsequent movement along this structure may have both controlled the deposition of the Glenfinnan and Loch Eil Divisions, and been responsible for tectonic sliding in the Loch Eil area.

#### 8.2.2.2. The Great Glen Fault

Since movement along major transcurrent faults is often preceded by an extensive history of ductile deformation (Watterson 1975), it was hoped that an E-W structural study across the Loch Eil Division migh reveal traces of an eastward increase in ductile deformation related to subsequent brittle displacement along the Great Glen Fault. However, there does not appear to be any systematic variation in the level of ductile deformation with respect to the Great Glen Fault, and movement along this structure, at least at the present tectonic level, seems to have occurred entirely along narrow zones of brittle deformation.

## REFERENCES

AFTALION, M. and VAN BREEMEN, O. 1980. U-Pb zircon, monazite and Rb-Sr whole rock systematics of granitic gneiss and psammitic to semi-pelitic host gneiss from Glenfinnan, northwestern Scotland. Contrib. Mineral. Petrol. 72, 97-98.

ALLEN, J.R.L. 1970. Sediments of the modern Niger delta: A summary and review. In: Morgan, J.P. (ed.), Deltaic Sedimentation, Modern and Ancient, 138-151. Soc. Econ. Palaeontologists and Mineralogists, Spec. Publ. No. 15.

ALLEN, J.R.L. and BANKS, N.L. 1972. An interpretation and analysis of recumbent-folded deformed cross-bedding. Sedimentology, 19, 257-283.

ANDERTON, R. 1976. Tidal shelf sedimentation: an example from the Scottish Dalradian. Sedimentology, 23, 429-458.

ANDERTON, R. 1982. Dalradian deposition and the late Precambrian-Cambrian history of the N. Atlantic region: a review of the early evolution of the Iapetus Ocean. J. Geol. Soc. Lond. 139, 423-434.

ASHWORTH, J.R. 1975. The sillimanite zones of the Huntley-Portsoy area in the northeast Dalradian, Scotland. Geol. Mag. 112, 113-36.

ASHWORTH, J.R. 1976. Petrogenesis of migmatites in the Huntly-Portsoy area, northeast Scotland. Mineralog. Mag. 40, 661-682.

BAILEY, E.B. and ANDERSON, E.M. 1925. The Geology of Staffa, Iona and Western Mull. Mem. Geol. Surv.

BAILEY, E.B. and MAUFE, H.B. 1960. The Geology of Ben Nevis and Glen Coe, and the Surrounding Country (Explanation of Sheet 53). Mem Geol. Surv. Scotland. H.M.S.O.

BELDERSON, R.L. and STRIDE, A.H. 1966. Tidal fashioning of a basal bed. Mar. Geol. 4, 237-257.

BENTLEY, A. 1970. Sedimentation studies in the Lower Greensand. A Report on the Sedimentology of the Aptian and Lower Albian strata near Leighton Buzzard, Bedfordshire. Unpubl. B.Sc. Thesis, University of Oxford.

BREWER, M.S., BROOK, M. and POWELL, D. 1979. Dating the tectono-metamorphic history of the south-western Moine, Scotland. In: Harris, A.L., Holland, C.H. and Leake, B.E. (eds.), The Caledonides of the British Isles - reviewed. Spec. Publ. Geol. Soc. Lond. No. 8, 129-138.

BROOK, M., BREWER, M.S. and POWELL, D. 1976. Grenville age for rocks in the Moine of north-western Scotland. Nature, London. 260. 515-517.

BROOK, M., POWELL, D. and BREWER, M.S. 1977. Grenville events in Moine rocks of the Northern Highlands, Scotland. J. Geol. Soc. Lond. 133, 489-496.

BROWN, R.L., DALZIEL, I.W.D. and JOHNSON, M.R.W. 1970. A review of the structure and stratigraphy of the Moinian of Ardgour, Moidart and Sunart-Argyll and Inverness-shire. Scott. J. Geol. 6, 309-335.

CLIFFORD, T.N. 1957. The stratigraphy and structure of part of the Kintail district of southern Ross-shire - its relationship to the Northern Highlands. Q. J. Geol. Soc. Lond. 113, 57-92.

COLEMAN, J.M. and WRIGHT, L.D. 1975. Modern river deltas: variability of processes and sand bodies. In: Broussard, M.L. (ed.), Deltas, Models for Exploration, 99-149. Houston Geol. Soc. Houston.

COLLINSON, J.D. 1970. Bedforms of the Tans River, Norway. Geogr. Annaler. 52-A, 31-36.

COLLINSON, J.D. 1978. Alluvial sediments. In: Reading, H.G. (ed.), Sedimentary environments and facies, 15-60. Blackwell Scientific Publications, Oxford.

CURRAY, J.R. 1964. Transgressions and regressions. In: Mills, R.L. (ed.), Papers in Marine Geology: Shepard Commemorative Volume, 175-203. MacMillan, New York.

DALZIEL, I.W.D. 1963. A structural study of the granite gneiss and associated Moine rocks between Loch Shiel and Loch Eil, Argyll and Inverness-shire. Unpubl. Ph.D. thesis, University of Edinburgh.

DALZIEL, I.W.D. 1966. A structural study of the granitic gneiss of western Ardgour, Argyll and Inverness-shire. Scott. J. Geol. 2, 125-152.

DEARNLEY, R. 1967. Metamorphism of minor intrusions associated with the Newer Granites of the Western Highlands of Scotland. Scott. J. Geol. 3, 449-457.

DEEGAN, C.E. 1971. The mode of origin of some late diagenetic sandstone concretions from the Scottish Carboniferous. Scott. J. Geol. 7, 357-365.

DRAKE, M.J. 1976. Plagioclase-melt equilibria. Geochim. et Cosmochim. Acta, 40, 457-466.

DREVER, H.I. 1940. The Geology of Ardgour, Argyllshire. Trans. Roy. Soc. Edin. 60, 141-170.

ELLIOTT, T. 1978. Deltas. In: Reading, H.G. (ed.), Sedimentary environments and facies, 97-142. Blackwell Scientific Publications, Oxford.

EYLES, V.A. and MACGREGOR, A.G. 1952. The Great Glen crush-belt. Geol. Mag. 89, 426-436.

FETTES, D.J. 1979. A metamorphic map of the British and Irish Caledonides. In: Harris, A.L., Holland, C.H. and Leake, B.E. (eds.), The Caledonides of the British Isles - reviewed. Spec. Publ. Geol. Soc. Lond. No. 8, 307-321.

FETTES, D.J. and MACDONALD, R. 1978. Glen Garry vein complex. Scott. J. Geol. 14, 335-358.

FLEUTY, M.J. 1964. The description of folds. Proc. Geol. Assoc. 75, 461-492.

GALLOWAY, W.E. 1975. Process framework for describing the morphologic and stratigraphic evolution of the deltaic depositional systems. In: Broussard, M.L. (ed.), Deltas, Models for Exploration, 87-98. Houston Geol. Soc. Houston.

GEIKIE, A. 1888. Report on the Recent Work of the Geological Survey in the North-West Highlands of Scotland. Q. J. Geol. Soc. Lond. 44, 378-441.

GOULD, D. 1966. Geochemical and mineralogical studies of the granitic gneiss and associated rocks of western Ardgour, Argyll. Unpubl. Ph.D. thesis, University of Edinburgh.

HARMS, J.C., SOUTHARD, J., SPEARING, D.L. and WALKER, R.G. 1975. Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Lecture Notes: Soc. econ. Paleont. Miner., Short Course 2, Dallas, Texas.

HARRIS, A.L., BALDWIN, C.T., BRADBURY, H.J., JOHNSON, H.D. and SMITH, R.A. 1978. Ensialic basin sedimentation: the Dalradian Supergroup. In: Bowes, D.R. and Leake, B.E. (eds.), Crustal evolution in northwestern Britain and adjacent regions. Geol. J. Spec. Issue No. 10, 115-38.

HARRIS, A.L. and RATHBONE, P. 1980. The Moine and Dalradian. In: Owen, T.R. (ed.), United Kingdom: Introduction to general geology and guides to excursions 002, 055, 093 and 151. Int. geol. Congr., 26th, Paris 1980.

HARRY, W.T. 1953. The composite granitic gneiss of Western Ardgour. Q. J. Geol. Soc. Lond. 109, 285-310.

HARTLEY, J. and LEEDAL, G.P. 1951. A Monchiquite Vent, Stob a 'Ghrianain,' Inverness-shire. Geol. Mag. 88, 140-144.

HASELOCK, P.J. 1982. The geology of the Corrieyaireck Pass area, Inverness-shire. Unpubl. Ph.D. thesis, University of Keele.

HAYES, J.B. 1964. Geodes and concretions from the Mississippian Warsaw Formation, Keokuk region, Iowa, Illinois, Missouri. J. Sedim. Petrol. 34,123-133.

HIGGINS, M.W. 1971. Cataclastic rocks. Geol. Surv. Prof. Paper 687. U.S. Govt. Printing Office, Washington.

HOBBS, B.E., MEANS, W.D. and WILLIAMS, P.E. 1976. An Outline of Structural Geology. John Wiley and Sons, Inc. New York.

HOLLAND, C.H. et al. 1978. A guide to stratigraphical procedure. Geol. Soc. Lond. Spec. Rep. No. 10.

HOWARD, J.D. and REINECK, H.E. 1972. Georgia coastal region, Sapelo Island, U.S.A.: Sedimentology and biology. IV. Physical and biogenic sedimentary structures of the nearshore Shelf. Senckenbergiana. Marit. 4, 81-123.

HUDLESTON, P.J. 1973. Fold morphology and some geometrical implications of theories of fold development. Tectonophysics, 16, 1-46.

HUTTON, D.H.W. 1981. Tectonic slides in the Caledonides. In: McClay, K.R. and Price, N.J. (eds.), Thrust and Nappe Tectonics. Spec. Publ. Geol. Soc. Lond. No. 9, 261-265.

JOCELYN, J. and PIDGEON, R.T. 1974. Examples of twinning and parallel growth in zircons from some Precambrian granites and gneisses. Mineral. Mag. 39,587-594.

JOHANNES, W. and GUPTA, L.N. 1982. Origin and Evolution of a Migmatite. Contrib. Mineral. Petrol. 79, 114-123.

JOHNSON, H.D. 1977. Sedimentation and water escape structures in some late Precambrian shallow marine sandstones from Finnmark, North Norway. Sedimentology, 24, 389-411.

JOHNSON, H.D. 1978. Shallow siliciclastic seas. In: Reading, H.G. (ed.), Sedimentary environments and facies, 207-258. Blackwells Scientific Publications, Oxford.

JOHNSON, M.R.W., SANDERSON, D.J. and SOPER, N.J. 1979. Deformation in the Caledonides of England, Ireland and Scotland. In: Harris, A.L., Holland, C.H. and Leake, B.E. (eds.), The Caledonides of the British Isles - reviewed. Spec. Publ. Geol. Soc. Lond. No. 8, 165-186.

JOHNSTONE, G.S. 1975. The Moine Succession. In: Harris, A.L., Holland, C.H. and Leake, B.E. (eds.), A Correlation of the Precambrian rocks of the British Isles. Spec. Publ. Geol. Soc. Lond. No. 6, 30-42.

JOHNSTONE, G.S., SMITH, D.I. and HARRIS, A.L. 1969. Moinian Assemblage of Scotland. Mem. Am. Assoc. Pet. Geol. 12, 159-80.

KENNEDY, W.Q. 1946. The Great Glen Fault. Q. J. Geol. Soc. Lond. 102, 41-76.

KENNEDY, W.Q. 1949. Zones of progressive metamorphism in the Moine schists of the Western Highlands of Scotland. Geol. Mag. 86, 43-56.

KENYON, N.H. 1970. Sand ribbons of European tidal Seas. Mar. Geol. 9, 25-39.

KENYON, N.H. and STRIDE, A.H. 1970. The tide swept continental shelf sediments between the Shetland Isles and France. Sedimentology 14, 159-173.

KLEIN, G. Dev. 1970. Tidal origin of a Precambrian quartzite - the Lower Fine-grained Quartzite (Middle Dalradian) of Islay, Scotland. J. Sedim. Petrol. 40, 973-985.

KLEIN, G. Dev. 1977. Clastic tidal facies. Cepco, Illinois.

KNILL, J.L. and KNILL, D.C. 1958. Some discordant fold structures from the Dalradian of Craignish, Argyll and Rosguile, Co. Donegal. Geol. Mag., 95, 497-510.

KRUMBEIN, W.C. and GARRELS, R.M. 1952. Origin and classification of chemical sediments in terms of pH and oxidation-reduction potentials. J. Geol. 60, 1-33.

KULM, L.D., ROUSCH, R.C., HARLETT, J.C., NEUDECK, R.H., CHAMBERS, D.M. and RUNGE, E.J. 1975. Oregon continental shelf sedimentation: interrelationships of facies distribution and sedimentary processes. J. Geol. 83, 145-176.

LAMBERT, R. St. J. 1969. Isotopic studies relating to the Pre-Cambrian history of the Moinian of Scotland. Proc. Geol. Soc. Lond. 1652, 243-245.

LAMBERT, R. St. J., WINCHESTER, J.A. and HOLLAND, J.G. 1979. Time, space and intensity relationships of the Precambrian and Lower Palaeozoic metamorphisms of the Scottish Highlands. In: Harris, A.L., Holland, C.H. and Leake, B.E. (eds.), The Caledonides of the British Isles - reviewed. Spec. Publ. Geol. Soc. Lond. No. 8, 363-7.

LEEDAL, G.P. 1952. The Cluanie igneous intrusion, Inverness-shire and Ross-shire. Q. J. Geol. Soc. Lond. 108, 35-62.

LEVELL, B.K. 1980. A late Precambrian tidal shelf deposit, the Lower Sandfjord Formation, Finnmark, North Norway. Sedimentology, 27, 539-557.

LIPPMANN, F. 1955. Ton, Geoden und Minerale des Barreme von Hoheneggelsen. Geol. Rundschau, 43, 475-503.

LOWE, D.R. 1975. Water escape structures in coarse-grained sediments. Sedimentology, 22, 157-204.

LOWE, D.R. 1976. Subaqueous liquefied and fluidized sediment flows and their deposits. Sedimentology, 23, 285-308.

MACGREGOR, A.G. 1948. Resemblances between Moine and "Sub-Moine" metamorphic sediments in the Western Highlands of Scotland. Geol. Mag. 85, 265-75.

MACKNIGHT, T. 1811. On the Mineralogy and Local Scenery of certain Districts in the Highlands of Scotland. Mem. Wernerian Nat. Hist. Soc. 1, 274-357.

McCAVE, I.N. 1972. Transport and escape of fine-grained sediment from shelf areas. In: Swift, D.J.P., Duane, D.B. and Pilkey, O.H. (eds.), Shelf Sediment Transport: Process and Pattern, 225-248. Dowden, Hutchinson and Ross Inc., Stroudsberg.

MEHNERT, K.R. 1968. Migmatites and the origin of granitic rocks. Amsterdam, Elsevier.

MENDUM, J.R. 1979. Caledonian thrusting in NW Scotland. In: Harris, A.L., Holland, C.H. and Leake, B.E. (eds.), The Caledonides of the British Isles - reviewed. Spec. Publ. Geol. Soc. Lond. No. 8, 291-297.

MERCY, E.L.P. 1963. The geochemistry of some Caledonian granitic and metasedimentary rocks. In: Johnson, M.R.W. and Stewart, F.H. (eds.), The British Caledonides, 189-215. Edinburgh.

MIDDLETON, G.V. 1970. Experimental Studies related to problems of Flysch sedimentation. In: Lajoie, J. (ed.), Flysch Sedimentology in North America, 253-272. Geol. Assoc. Can. Spec. Pap. No. 7.

MILLIMAN, J.D. and EMERY, K.O. 1968. Sea levels during the past 35,000 years. Science, 162, 1121-1123.

MURCHISON, R.I. and GEIKIE, A. 1861. On the Altered Rocks of the Western Islands of Scotland, and the North-western and Central Highlands. Q. J. Geol. Soc. Lond. 17,171-240.

PANKHURST, R.J. and PIDGEON, R.T. 1976. Inherited isotope systems and the source region pre-history of early Caledonian granites in the Dalradian Series of Scotland. Earth and Planetary Science Letters, 31,55-68.

PANTIN, H.M. 1958. Rate of formation of a diagenetic calcareous concretion. J. Sedim. Petrol. 28, 366-371.

PEACOCK, J.D. 1977. Metagabbros in granitic gneiss, Inverness-shire, and their significance in the structural history of the Moines. Rep. Inst. Geol. Sci., No. 77/20.

PHEMISTER, J. 1948. The Northern Highlands. British Regional Geology, Geol. Surv.

PIASECKI, M.A.J. 1980. New light on the Moine rocks of the Central Highlands of Scotland. J. Geol. Soc. Lond. 137, 41-59.

PIASECKI, M.A.J., VAN BREEMEN, O. and WRIGHT, A.E. 1981. Late Pre-Cambrian geology of Scotland, England and Wales. Canad. Soc. Pet. Geol. Mem. 7, 57-94.

PIDGEON, R.T., and AFTALION, M. 1978. Cogenetic and inherited zircon U-Pb systems in Palaeozoic granites of Scotland and England. In: Bowes, D.R., and Leake, B.E. (eds.), Crustal evolution in northwest Britain and adjacent regions. Geol. J. Spec. Issue No. 10, 183-200.

POLDERVAART, A. 1950. Statistical Studies of Zircon as a Criterion in Granitisation. Nature, 165, 574-575.

POLDERVAART, A. 1956. Zircon in rocks: 2. Igneous rocks. Amer. J. Sci. 254, 521-554.

POWELL, D. 1964. The stratigraphical succession of the Moine schists around Lochailort (Inverness-shire) and its regional significance. Proc. Geol. Assoc. 75, 223-246.

POWELL, D. 1974. Stratigraphy and structure of the western Moine and the problem of Moine orogenesis. J. Geol. Soc. Lond. 130, 575-593.

POWELL, D., BAIRD, A.W., CHARNLEY, N.R. and 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. Lond. 136, 661-673.

RAAF, J.F.M. De and BOERSMA, J.R. 1971. Tidal deposits and their sedimentary structures. Geologie Mijnb. 50, 479-504.

RAAF, J.F.M. De, BOERSMA, J.R. and GELDER, A. Van. 1977. Wave-generated structures and sequences from a shallow marine succession, Lower Carboniferous, County Cork, Ireland. Sedimentology, 24, 451-483.

RAISWELL, R. 1971. The growth of Cambrian and Liassic concretions. Sedimentology, 17, 147-171.

RAMBERG, H. 1952. The origin of metamorphic and metasomatic rocks. Chicago, Univ. Chicago Press.

RAMBERG, H. 1963. Strain distribution and geometry of folds. Bull. Geol. Inst. Univ. Upps. 42, 1-20.

RAMBERG, H. 1964. Selective buckling of composite layers with contrasted rheological properties. A theory for simultaneous formation of several orders of folds. Tectonophysics, 1, 307-341.

RAMSAY, J.G. 1961. The effects of folding upon the orientation of sedimentary structures. J. Geol. 69, 84-100.

RAMSAY, J.G. 1963. Structure and metamorphism of the Moine and Lewisian rocks of the North West Caledonides. In: Johnson, M.R.W. and Stewart, F.H. (eds.), The British Caledonides 143-175. Oliver and Boyd, Edinburgh.

RAMSAY, J.G. 1967. Folding and fracturing of rocks. McGraw-Hill Book Co., New York.

RAMSAY, D.M. and STURT, B.A. 1973. An analysis of noncylindrical and incongruous fold pattern from the Eo-Cambrian rocks of Söröy, Northern Norway. Tectonophysics, 18, 81-121.

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, University of Liverpool.

READING, H.G. 1978. Facies. In: Reading, H.G. (ed.), Sedimentary environments and facies, 4-14. Blackwell Scientific Publications, Oxford.

REINECK, H.E. and SINGH, I.B. 1973. Depositional sedimentary environments - with reference to terrigenous clastics. Springer-Verlag, Berlin.

RETTGER, R.E. 1935. Experiments on soft-rock deformation. J. Geol. 19, 271-292.

RHODES, S. and GAYER, R.A. 1977. Non-cylindrical folds, linear structures in the X direction and mylonite developed during translation of the Caledonian Kalak Nappe Complex of Finnmark. Geol. Mag. 114, 329-341.

RICHEY, J.E. 1931. West Highland District. Summ. Progr. Geol. Surv. for 1930, part 1, 63-4.

RICHEY, J.E. and KENNEDY, W.Q. 1939. The Moine and Sub-Moine Series of Morar, Inverness-shire. Bull. Geol. Surv. G.B., 2, 26-45.

ROBERTS, J.L. 1974. The structure of the Dalradian rocks in the SW Highlands of Scotland. J. Geol. Soc. Lond. 130, 93-124.

SEED, H.B. 1968. Landslides during earthquakes due to soil liquefaction. Proc. Am. Soc. Civil Eng., J. Soil Mech. Found. Div. 94(SM5), 1055-1122.

SHROCK, R.R. 1948. Sequence in layered rocks. McGraw-Hill Book Co., New York.

SINGH, I.B. 1969. Primary sedimentary structures in Precambrian quartzites of Telemark, southern Norway, and their environmental significance. Norsk. Geol. Tidsskr. 49, 1-31.

SMITH, D.I. 1979. Caledonian minor intrusions of the N. Highlands of Scotland. In: Harris, A.L., Holland, C.H. and Leake, B.E. (eds.), The Caledonides of the British Isles - reviewed. Spec. Publ. Geol. Soc. Lond. No. 8, 683-697.

SOPER, N.J. 1960. A note on current-bedded Moines near Garve, Wester Ross. Geol. Mag. 6, 504-508.

SOPER, N.J. and BARBER, A.J. 1982. A model for the deep structure of the Moine thrust zone. J. Geol. Soc. Lond. 139, 127-138.

SPRY, A. 1969. Metamorphic textures. Pergammon Press, Oxford.

STANLEY, D.J. 1974. Pebbly mud transport in the head of Wilmington Canyon. Marine Geol. 16, M1-M8.

STEWART, A.D. 1982. Late Proterozoic rifting in NW Scotland: the genesis of the 'Torridonian'. J. Geol. Soc. Lond. 139, 415-422.

STOKER, M.S. 1980. The Geology of Eastern Ardgour. Unpubl. Ph.D. thesis, University of Liverpool.

SWETT, K., KLEIN, G. Dev. and SMIT, D.E. 1971. A Cambrian tidal sand body, the Eriboll Sandstone of northwest Scotland: an ancient-recent analog. J. Geol. 79, 400-415.

SWIFT, D.J.P. 1976. Continental Shelf Sedimentation. In: Stanley, D.J. and Swift, D.J.P. (eds.), Marine Sediment Transport and Environmental Management, 311-350. John Wiley, New York.

TANNER, P.W.G. 1976. Progressive regional metamorphism of thin calcareous bands from the Moinian rocks of N.W. Scotland. J. Petrol. 17, 100-34.

TANNER, P.W.G., JOHNSTONE, G.S., SMITH, D.L. and HARRIS, A.L. 1970. Moine stratigraphy and the problem of the Central Ross-shire inliers. Geol. Soc. Am. Bull. 81, 299-306.

THOMAS, P.R. 1980. The structure and stratigraphy of the Moine rocks N. of the Schiehallion complex, Scotland. J. Geol. Soc. Lond. 137, 469-482.

TURNER, F.J. 1981. Metamorphic petrology (2nd ed.). McGraw-Hill Book Co., New York.

VAN BREEMEN, O., PIDGEON, R.T. and JOHNSON, M.R.W. 1974. Precambrian and Palaeozoic pegmatites in the Moines of northern Scotland. J. Geol. Soc. Lond. 130, 493-507.

VAN BREEMEN, O., HALLIDAY, A.N., JOHNSON, M.R.W. and BOWES, D.R. 1978. Crustal additions in late Precambrian times. In: Bowes, D.R. and Leake, B.E. (eds.), Crustal evolution in northwestern Britain and adjacent regions. Geol. J. Spec. Issue No. 10, 81-106.

VAN BREEMEN, O., AFTALION, M., PANKHURST, R.J. and RICHARDSON, S.W. 1979. Age of the Glen Dessary Syenite, Inverness-shire: diachronous Palaeozoic metamorphism across the Great Glen. Scott. J. Geol. 15, 49-62.

VOLL, G. 1960. New work on petrofabrics. Liverp. Manch. Geol. J., 2, 503-567.

VON PLATEN, H. 1965. Experimental anatexis and genesis of megmatites. In: Pitcher, W.S., and Flinn, G.W. (eds.), Controls of Metamorphism. Oliver and Boyd, Edinburgh, 203-218.

WALKER, R.G. 1979. Turbidites and Associated Coarse Clastic Deposits. In: Walker, R.G. (ed.), Facies Models, 91-104. Geol. Assoc. Canada.

WALKER, R.G. and CANT, D.J. 1979. Sandy Fluvial Systems. In: Facies Models (Ed. by R.G. Walker), 23-32. Geol. Assoc. Canada.

WATTERSON, J. 1975. Mechanism for the persistence of tectonic lineaments. Nature, London, 253, 520-522.

WENK, E. and KELLER, F. Isograde in Amphibolitserien der Zentralalpen. Schweiz. Mineral. Petrogr. Mitteil. 49, 157-198.

WENTWORTH, C.M. 1967. Dish structure, a primary sedimentary structure in coarse turbidites (abstract). Am. Assoc. Pet. Geol. Bull. 51,485.

WHITE, A.J.R. 1966. Genesis of migmatites from the Palmer region of South Australia. Chem. Geology, 1, 165-200.

WHITTEN, E.H.T. 1966. Structural geology of folded rocks. Rand McNally and Co. U.S.A.

WHITTLES, K.H. 1981. The geology and geochemistry of the area west of Loch Killin, Inverness-shire. Unpubl. Ph.D. thesis, University of Keele.

WILSON, C.J.L. 1973. The prograde microfabric in a deformed quartzite sequence, Mount Isa, Australia. Tectonophysics, 19, 39-81.

WILSON, G., WATSON, J. and SUTTON, J. 1953. Current-bedding in the Moine Series of North-Western Scotland. Geol. Mag. 90, 377-386.

WINCHESTER, J.A. 1970. The Geology of Fannich Forest. Unpubl. D.Phil. thesis, University of Oxford.

WINCHESTER, J.A. 1972. The petrology of Moinian calc-silicate gneisses from Fannich Forest, and their significance as indicators of metamorphic grade. J. Petrol. 13, 405-24.

WINCHESTER, J.A. 1973. Pattern of regional metamorphism suggests a sinistral displacement of 160km along the Great Glen Fault. Nature, Phys. Sci. Lond. 246, 81-84.

WINCHESTER, J.A. 1974. The zonal pattern of regional metamorphism in the Scottish Caledonides. J. Geol. Soc. Lond. 130, 509-524.

WINCHESTER, J.A. 1975. Epidotic calc-silicates of Arnipol Type - a widely distributed minor sedimentary facies of the Moine assemblage, Scotland. Geol. Mag. 2, 175-181.

WINCHESTER, J.A. 1976. Different Moinian amphibolite suites in northern Ross-shire. Scott. J. Geol. 12, 187-204.

WINCHESTER, J.A., LAMBERT, R. St. J. and HOLLAND, J.G. 1981. Geochemistry of the western part of the Moinian assemblage. Scott. J. Geol. 17, 281-294.

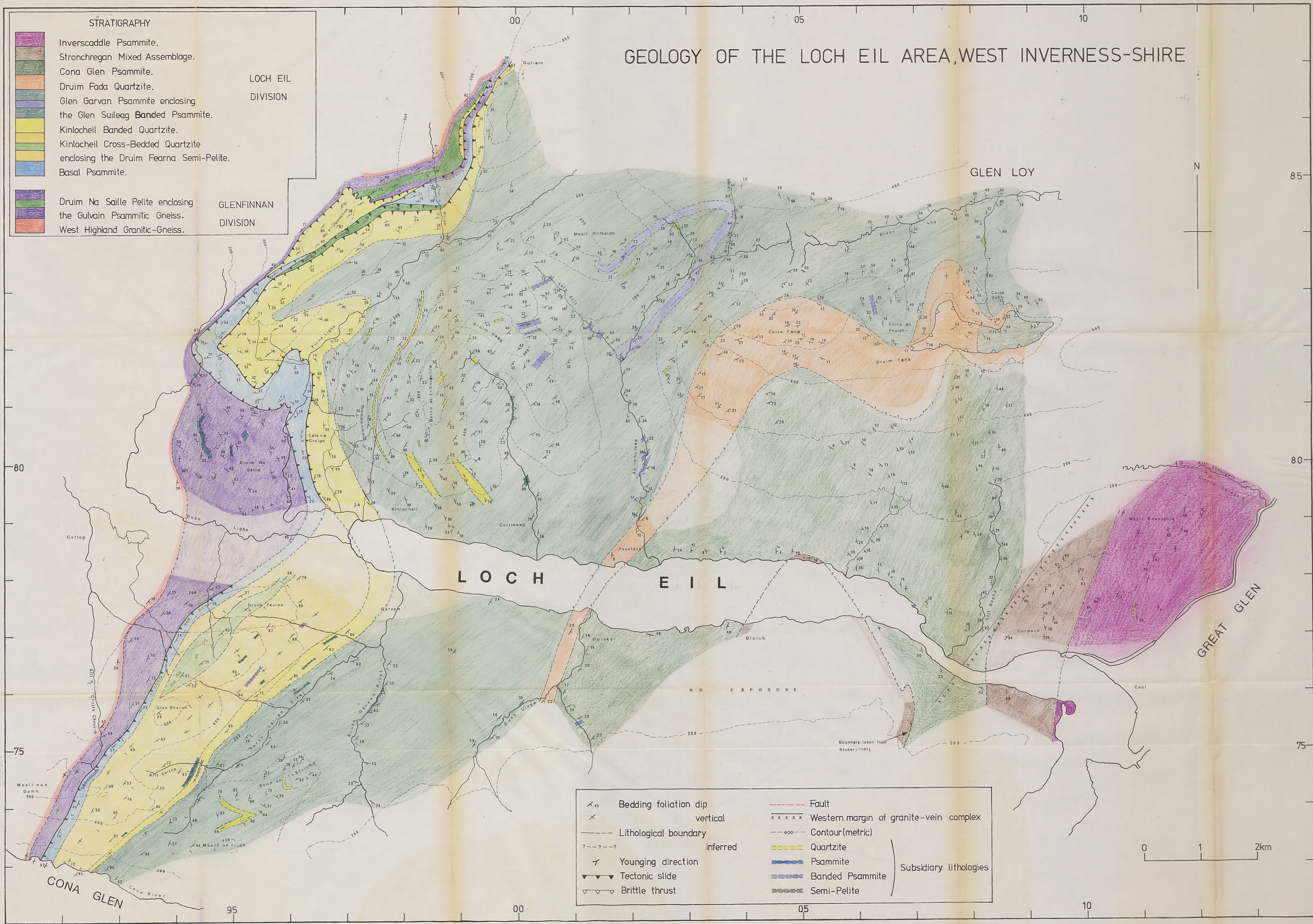
WINKLER, H.G.F. 1979. Petrogenesis of metamorphic rocks (4th ed.). New York, Springer-Verlag.

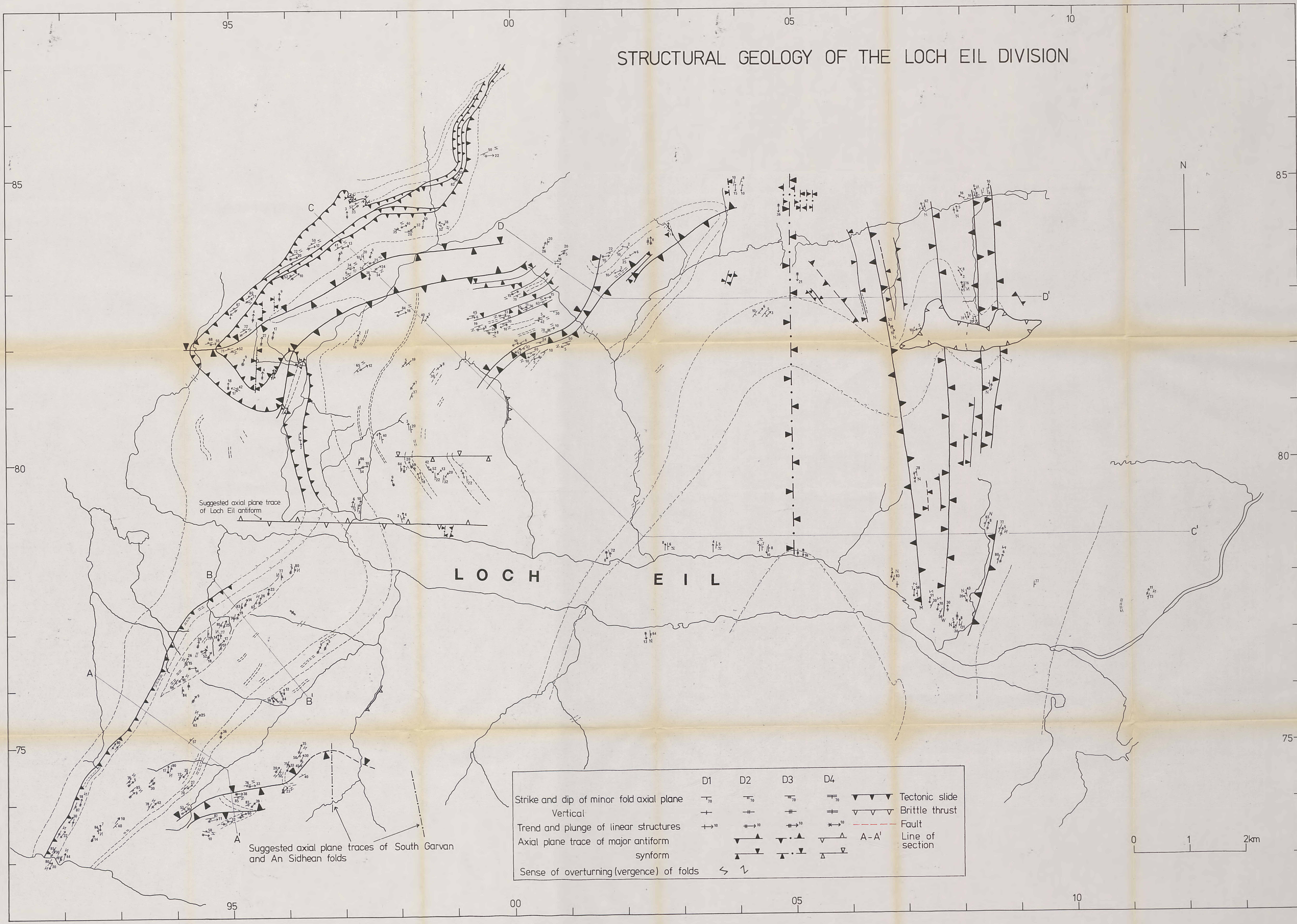
WISEMAN, J.D.H. 1934. The central and southwest Highland epidiorites. Q. J. Geol. Soc. Lond. 90, 354-417.

WRIGHT, L.D. 1977. Sediment transport and deposition at river mouths: a synthesis. Bull. Geol. Soc. Am. 88, 857-868.

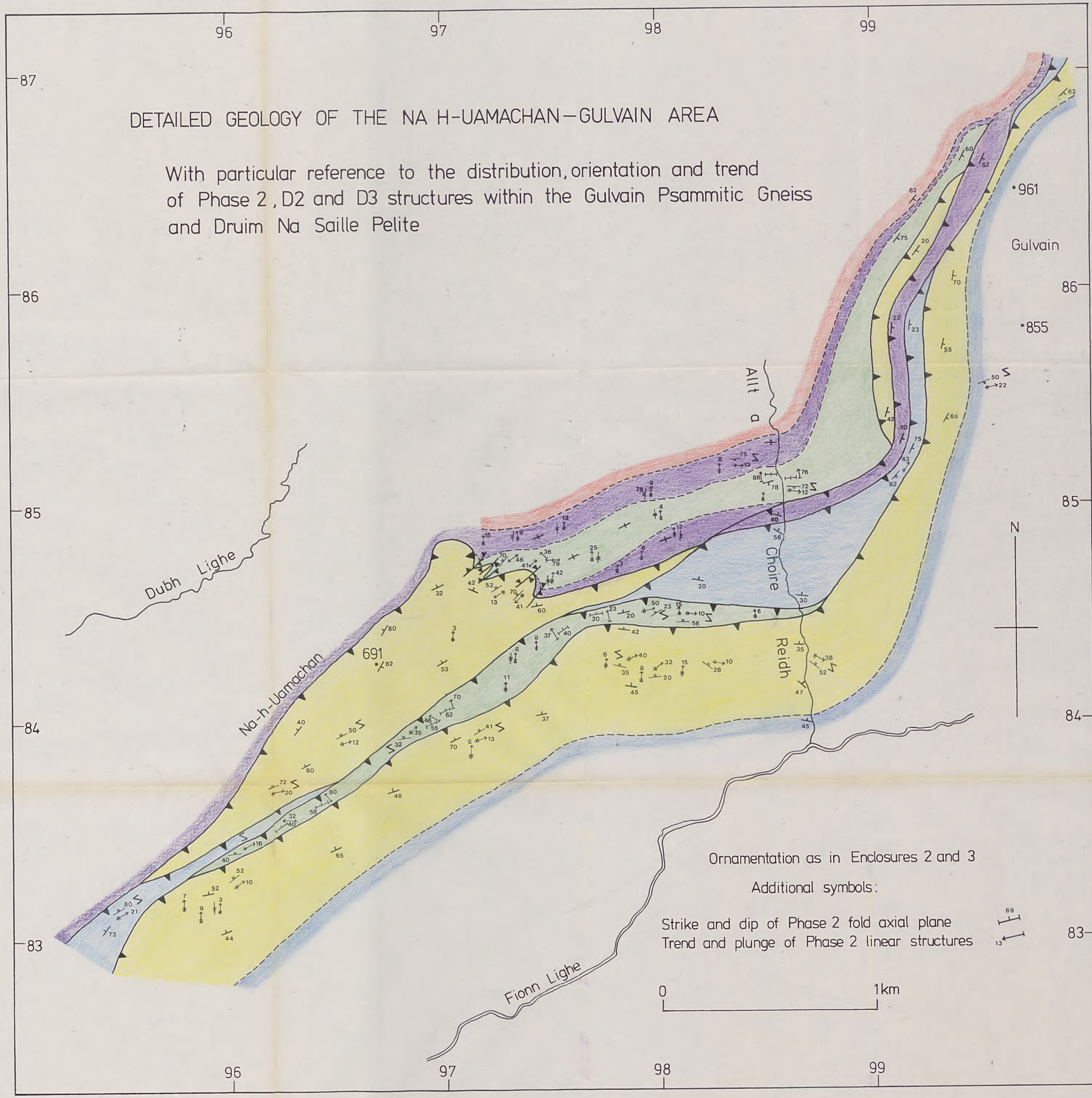
YARDLEY, B.W.D. 1978. Genesis of the Skagit Gneiss migmatites, Washington, and the distinction between possible mechanisms of migmatisation. Bull. Geol. Soc. Am. 89, 941-951.

SYNOPSIS OF THE GEOLOGY OF THE LOCH EIL AREA
STRATIGRAPHY
Inverscaddle Psammite
Stronchreggan Mixed Assemblage
Cona Glen Psammite
Druim Fada Quartzite
Glen Garvan Psammite enclosing
the Glen Suileag Banded Psammite
Kinlocheil Banded Quartzite
Kinlocheil Cross-Bedded Quartzite
enclosing the Druim Fearna Semi-Pelite
Basal Psammite
LOCH EIL
DIVISION
Druim Na Saille Pelite enclosing
the Gulvain Psammitic Gneiss
West Highland Granitic-Gneiss
GLENFINNAN
DIVISION
GLEN LOY
LOCH EIL
GREAT GLEN
CONA GLEN
N
Lithological boundary
Tectonic slide
Brittle thrust
Fault
Axial plane trace of D2 fold (antiform)
D3 (synform)
0
1
2km
00
05
10
95
85
80
75


GEOLOGY OF THE LOCH EIL AREA, WEST INVERNESS-SHIRE
STRATIGRAPHY
Inverscaddle Psammite.
Stronchregan Mixed Assemblage.
Cona Glen Psammite.
Druim Fada Quartzite.
Glen Garvan Psammite enclosing
the Glen Suileag Banded Psammite.
Kinlocheil Banded Quartzite.
Kinlocheil Cross-Bedded Quartzite
enclosing the Druim Fearna Semi-Pelite.
Basal Psammite.
LOCH EIL DIVISION
Druim Na Saille Pelite enclosing
the Gulvain Psammitic Gneiss.
West Highland Granitic-Gneiss.
GLENFINNAN DIVISION
Bedding foliation dip
vertical
Lithological boundary
inferred
Younging direction
Tectonic slide
Brittle thrust
Fault
Western margin of granite-vein complex
Contour (metric)
Quartzite
Psammite
Banded Psammite
Semi-Pelite
Subsidiary lithologies
0
1
2km
N
LOCH EIL
GLEN LOY
GREAT GLEN
CONA GLEN
NO EXPOSURE
Gulvain
Kinlocheil
Corriebeg
Garvan
Callop
Druim Fearna
Druim Fada
Corpach
Caol
Blaich
Meall Bhanabhie
Boundary taken from Stoker (1981)
00
05
10
95
80
85
75


STRUCTURAL GEOLOGY OF THE LOCH EIL DIVISION
95
00
05
10
85
80
75
N
LOCH EIL
Suggested axial plane trace of Loch Eil antiform
Suggested axial plane traces of South Garvan and An Sidhean folds
A
A'
B
B'
C
C'
D
D'
D1
D2
D3
D4
Strike and dip of minor fold axial plane
Vertical
Trend and plunge of linear structures
Axial plane trace of major antiform
synform
Sense of overturning (vergence) of folds
Tectonic slide
Brittle thrust
Fault
A-A'
Line of section
0
1
2km

# CROSS - SECTIONS


DETAILED GEOLOGY OF THE NA H-UAMACHAN—GULVAIN AREA
With particular reference to the distribution, orientation and trend of Phase 2, D2 and D3 structures within the Gulvain Psammitic Gneiss and Druim Na Saille Pelite
Gulvain
Allt a Choire Reidh
Dubh Lighe
Na-h-Uamachan
Fionn Lighe
Ornamentation as in Enclosures 2 and 3
Additional symbols:
Strike and dip of Phase 2 fold axial plane
Trend and plunge of Phase 2 linear structures
0
1km
N