QUATERNARY GEOMORPHOLOGY OF THE ASSYNT AREA, N.W. SCOTLAND

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DECLARATION

In compliance with regulation 2.4.15 of the University's Regulations for Postgraduate Study, I hereby declare that this thesis has been composed by myself from my own work.

Timothy John Lawson

DEDICATION

My paternal grandmother, Florence Alice Lawson, a kind and good lady, died on 8 March 1983 and therefore did not see this research project finished. Although she did not understand its content, she understood the importance of this thesis to me and reinforced my will to complete it. This thesis is therefore dedicated to her memory.

ABSTRACT

The main factors influencing the distribution of geomorphic features in the Assynt area are geology and the location of Loch Lomond Advance glaciers. A study of glacial striae and erratics has enabled the construction of a model for the development of the last ice sheet. Glaciers developed in the corries on the N and E side of the Assynt mountains, coalesced and thickened until an ice divide was established, ice overtopping the mountain ridge to flow both eastwards and westwards; the ice divide was always situated to the E side of the mountains. Even the highest parts of the study area were covered by ice to a considerable depth. An early deglaciation of the area is suggested by a radiocarbon date of 18,040 $\pm$ 240 years B.P.★ from fragments of reindeer antlers. Reconstruction of the seven Loch Lomond Advance glaciers that subsequently formed in the area has shown that the main snow-bearing airstreams came from the south and that the blowing of snow onto glaciers surfaces was a major factor in their development.

Glacial friction cracks are widespread in the Assynt area. A revision of nomenclature and a simplified classification is suggested. Attempts have been made to characterise certain Assynt friction crack forms, and orientation studies suggest that they are useful ways of establishing the former ice flow direction when striae or ice-moulded bedrock are absent, as long as a large number of them are measured. Their orientation is sometimes affected by weakenesses in the bedrock.

A study of the caves in the Cambrian dolomite of the area indicates that they originated phreatically, but subsequent lowering of the local water-table has tended to result in high-level,

abandoned passages, often choked with clastic deposits, and lower-level passages containing the active streamways. Clastic cave sediments are largely allochthonous, being derived from local glacial deposits. The dating of certain calcite speleothems has shown that many of the main elements of the subterranean drainage network were in existence prior to the last glaciation, and some parts may pre-date the penultimate interglacial. The Creag nan Uamh caves have yielded a unique Devensian and Flandrian fauna, and also evidence for the earliest recorded existence of man in Scotland, dated by radiocarbon to 10,080 $\pm$ 70 years B.P.

★ Subsequent radiocarbon dates from individual antlers and bones from the same stratigraphic horizon gave dates of 23,360 $^{+810}_{-740}$, 24,590 $^{+790}_{-720}$ and 8,300 $\pm$ 90 years B.P. This suggests that the date of 18,040 $\pm$ 240 years B.P. may be an average date derived from material of widely differing ages.

This is discussed in an appendix to the thesis.

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T.J.L.

Wellington College

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CONTENTS

LIST OF ILLUSTRATIONS

LIST OF TABLES

CHAPTER 1 : INTRODUCTION

The study area consists of a large part of the parish of Assynt, after which it has been named. It comprises some 450 km^2 of south-western Sutherland and part of Ross-shire. The limits of the field area follow physical features as much as possible, but in order to restrict the area to a manageable size certain of the boundaries follow O.S. grid lines and part of the A837 road NE of Lochinver. These boundaries are shown in Fig. 2.3.

The aim of this research was to elucidate the Quaternary geomorphology of an area of varied relief, believed to be situated somewhere near the position of the ice divide of part of the last ice sheet that covered the North-West Highlands of Scotland. It was hoped that a model of the detailed pattern of former ice flow could be constructed using various types of evidence. In this respect the Assynt area has the advantage that its geology is well mapped and, although structurally a complex region, it possesses only a few rock types, all of which are relatively easy to distinguish in the field. Furthermore, the lithological boundaries trend approximately at right angles to the anticipated E-W direction of ice flow. All these things were expected to aid a study of the distribution of glacial erratics in the area. Certain hard rock types, notably the Cambrian quartzite, favour the preservation of microerosional features that can also be used to determine the direction and trend of former ice flow. The limited access to the area has resulted in little previous work of this nature, which it was hoped would give scope for the current research project. The Assynt area also has the advantage of being covered by good quality aerial photographs and a selection of accurate maps at various scales (the O.S. 1:25,000 Second Series and 1:10,560 scale maps having photogrametrically-produced contours).

The presence of relatively thick strata of dolomitic limestone also gave a unique opportunity to study karst geomorphology on a scale unmatched elsewhere in Scotland, allowing the study of caves and their sedimentary deposits. It was hoped that such a study would provide palaeoenvironmental information and a chronological framework for the geomorphic evolution of the area.

The thesis is presented in thirteen chapters which can be subdivided into five main parts.

(i) The first two chapters introduce the study area to the reader.

(ii) The next five chapters consider the glacial history of the Assynt area. Chapter 3 describes and discusses the distribution of the main glacial and periglacial features. Chapters 4 and 5 consider the direction of former ice flow from studies of glacial erratics and striations respectively. Chapter 6 discusses the results of previous research in this part of NW Scotland, and the conclusions reached in chapters 4 and 5 are used to construct a model for the development of an ice sheet in the area. The glaciation of the field area is considered with regard to the present understanding of the Late Quaternary of Scotland. Chapter 7 attempts to make palaeoclimatic inferences from the reconstructed Loch Lomond Advance glaciers using the techniques developed by Sissons and his co-workers (*e.g.* Sissons & Sutherland 1976).

(iii) Glacial friction cracks are well represented in the Assynt area and are considered in a separate study (chapter 8). The classification and nomenclature is considered and redefined, and an attempt is made to characterise quantitatively certain of the friction crack forms.

(iv) Chapters 9 to 12 consider aspects of the geomorphology of the dolomite areas. Chapter 9 essentially describes the important elements of the drainage networks of the three main karstic drainage basins, and isolates two main topics for further research : the sedimentological analyses of the clastic cave deposits are studied in chapter 10, and the results of the $^{230}Th/^{234}U$ dating of certain speleothems are considered in chapter 11. Chapter 12 looks at the long-neglected Creag nan Uamh caves, reassessing the published and unpublished data and discussing the results of the author's own work in the caves in terms of the present understanding of the Late Quaternary history of the area.

(v) The final chapter (chapter 13) synthesises the main conclusions of this research and suggests where further work may be most beneficial.

CHAPTER 2 : THE FIELD AREA

2.1 GEOLOGY

2.1.1 Introduction

One of the most important constraints on geomorphological forms and processes is the influence of the local geology, both from the point of view of geological structure and lithological variation. In addition, a major part of this research concerns the study of the distribution of glacial erratic blocks. For these reasons it is necessary to outline briefly the geological succession and structure of the region.

Besides exposing a succession of some of the oldest rocks in the country, Assynt possesses some of the clearest evidence for the westward dislocation of rocks that comprises an important part of the early structural history of northern Britain. It is for these reasons that Assynt has become one of the classic areas of British geology. Much of the early survey work in the area was undertaken by B.N. Peach, J. Horne, C.T. Clough and L.W. Hinxman. The geological memoir for the North-West Highlands which deals with the Assynt area (Peach *et al*. 1907) is still the most detailed general reference work, an adequate testimony to the high standard of fieldwork of these early workers. Summaries of the geology of Assynt are to be found in Peach & Horne (1914), in the British Regional Geology series (Phemister 1960) and in the well known geological excursion guide (Macgregor & Phemister, 3rd. edition 1972), recently updated (Johnson & Parsons 1979).

2.1.2 Geological Succession

Fig. 2.1 schematically illustrates the major elements of the geological succession in the Assynt area. The terms for the Cambro-Ordovician follow Johnson & Parsons (1979) in using those given by Cowie *et al*. (1972) which differ slightly from those used on the Geological Survey special sheet of the Assynt district.

(a) Lewisian Group: This consists of gneisses of the Scourian complex (2,700 million years old), modified by the later Inverian and Laxfordian metamorphic events, and traversed by pre-Torridonian dykes trending in two distinct directions (WNW - ESE and E - W). These rocks have been affected by pre-Torridonian shear belts.

(b) Prolonged subaerial denudation of the Lewisian land surface produced a ground surface of considerable relief. On this surface sandstones and grits of the Torridonian Group were laid down from the west, under semi-arid conditions in lakes and as alluvial fans (800 million years ago).

(c) Gentle folding of these Pre-Cambrian strata was followed by extensive marine erosion that revealed the underlying Lewisian rocks in many places. Upon this almost level surface the earliest Cambrian sediments were deposited in an ever-deepening sea. The oldest rocks of this age are the 'False-bedded' Quartzite, pink or more often white in appearance, overlain by quartzites known as 'Pipe Rock' from the lithified sediment-infilled burrows of *Scolithus* and *Monocraterion* that they contain.

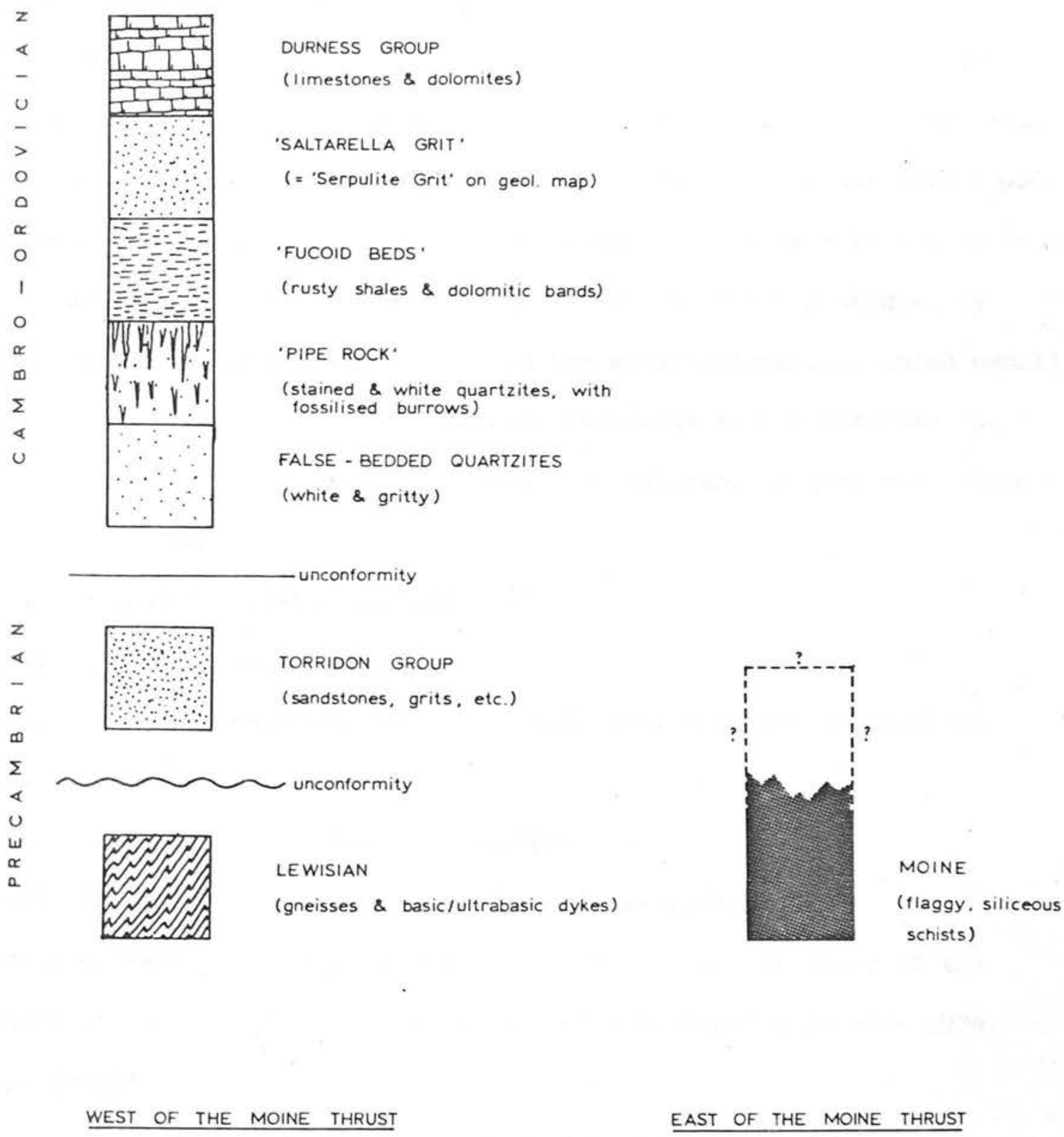
(d) The Cambrian quartzites were succeeded by fine-grained calcareous muds which now form the rust-coloured shales and dolomitic bands of the 'Fucoid' Beds.

(e) Deposited conformably on the 'Fucoid' Beds is the Salterella Grit, consisting of quartzite overlain by a dolomitic grit containing fossils

Fig. 2.1 Schematic section of the geological succession in the Assynt area.

(diagram overleaf)

THE GEOLOGICAL SUCCESSION IN THE ASSYNT DISTRICT



of the organism Salterella.

(f) The uppermost strata comprise the limestones, dolomitic limestones and dolomites of the Durness Group, deposited in clear waters. These lithologies are subdivided into the grey dolomites and limestones of the Grudaidh Formation, overlain by the white dolomites and limestones, with intermittent chert layers, of the Eilean Dubh Formation, and topped by dolomites of the Sailmhor Formation.

(g) The Cambrian and earlier rocks of the Assynt area are cut by a suite of alkaline sills and dykes, and two larger intrusive complexes, the Loch Borralan and Loch Ailsh masses. Together with the Loch Loyal intrusions farther north, these are the only extensive alkaline igneous intrusions in the U.K., and hence have been intensively studied by geologists. Sabine (1953) described the minor intrusives, which usually take the form of sills in the Torridonian rocks and in the Cambrian succession, with a few dykes cutting the Lewisian. Sabine recognised six main types :

i) grorudites (aegirine-felsites)

ii) Canisp porphyry)
)
iii) horneblende-porphyries) varieties of quartz-microsyenites
)
iv) nordmarkites)

v) vosgesites (horneblende-lamprophyres)

vi) ledmorites and borolanites (nepheline-syenites)

These different suites of rocks are limited to one or other of the dislocated nappes that are described below (Johnson & Parsons 1979, pp. 20-21).

The major intrusions have been investigated by various workers (e.g. Shand 1910, 1939; Phemister 1926). Initial views that both intrusions were stratified laccoliths have been revised by Woolley's (1970) work on the Loch Borralan Complex. Woolley has shown that the intrusion can be divided into two suites : the earlier suite was emplaced in the rocks of the Ben More nappe, and the later suite punched through both the earlier suite and the Ben More Thrust plane (section 2.1.3).

(h) To the east of the Moine Thrust are the crystalline quartzo-felspathic schists, with subordinate mica-schist beds, that are collectively know as the Moine schists. The age of these rocks is still uncertain, but it is generally accepted that considerable parts, if not all, of the Moine schists were deposited before 1,000 million years ago. Hence the Moine schists are of Pre-Cambrian age, and at least in part pre-date the Torridonian Group.

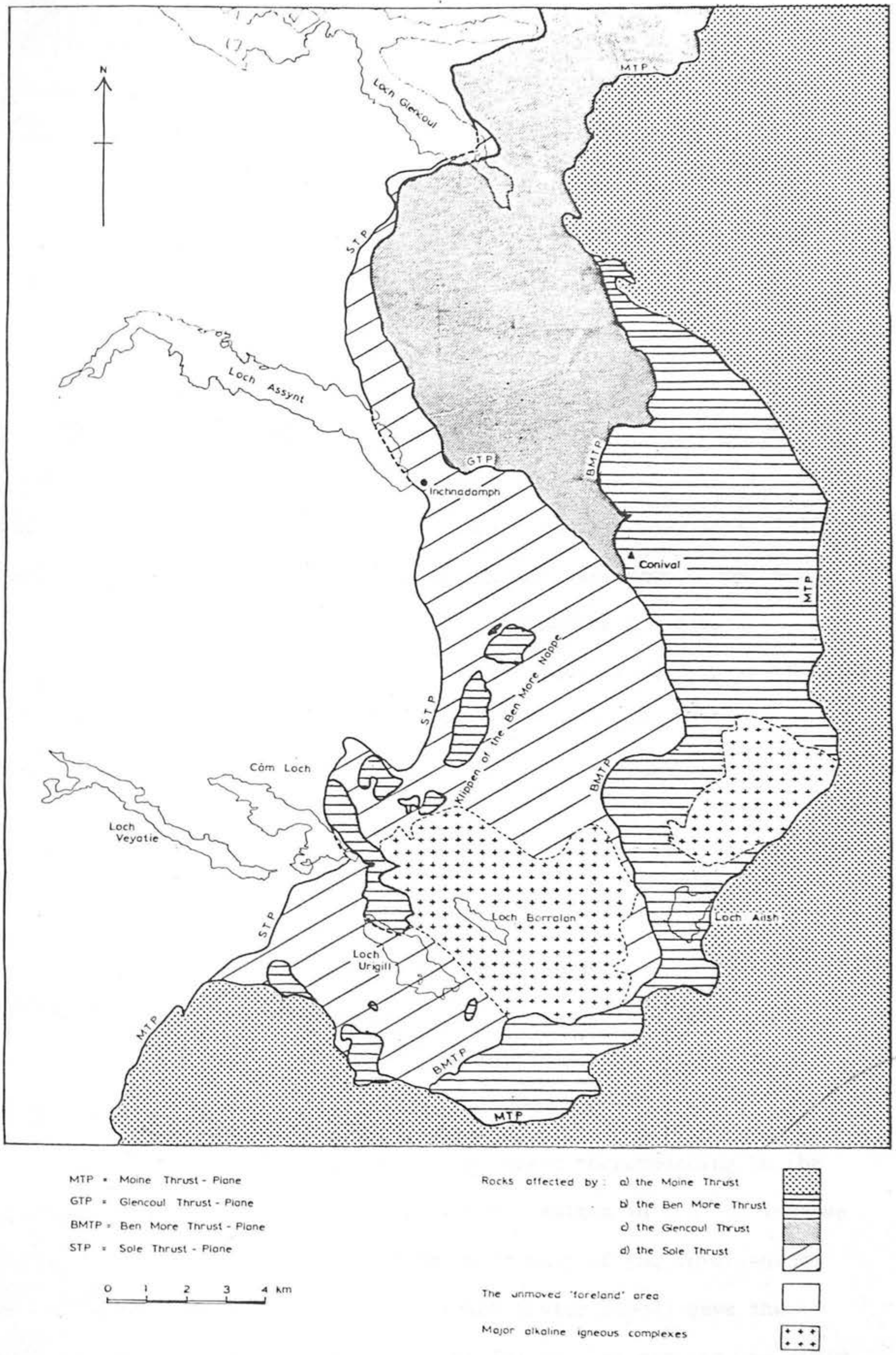
2.1.3 Geological Structure

The geological map of the Assynt district shows a complicated outcrop pattern. This is due to the presence of large, horizontally-directed dislocations or thrusts, along which masses of rock have been displaced westwards. There are four major lines of dislocation named, from west to east, the Sole Thrust, the Glencoul Thrust, the Ben More Thrust, and the Moine Thrust (Fig. 2.2). Minor lines of dislocation occur between the major thrusts; perhaps the most striking example of this is the imbricate structure formed by the numerous high-angle reverse faults above or below major low-angle thrusts running through the Cambrian dolomite area east and south-east of Inchnadamph.

The large embayment in the line of the Moine Thrust, apparent in Fig. 2.2, is the result of a broad upwarping of the Moine Thrust plane. Thus between the Moine Thrust and the Sole Thrust in this area there is a large lenticular mass of thrust material. The Moine nappe possibly extended far to the west, over the foreland area, as evidenced by a small outlier of Moine rock near Durness, farther north, which is about 10 km to the west of the main Moine outcrop. The presence of an

Fig. 2.2 Geological structure in the vicinity of the 'Assynt window', part of the Moine Thrust belt.

(diagram overleaf)


N
MTP
Loch Glencoul
STP
Loch Assynt
GTP
BMTP
Inchnadamph
Conival
MTP
STP
Klippen of the Ben More Nappe
Cam Loch
Loch Veyatie
BMTP
STP
Loch Borralan
Loch Ailsh
Loch Urigill
MTP
BMTP
MTP
MTP = Moine Thrust - Plane
GTP = Glencoul Thrust - Plane
BMTP = Ben More Thrust - Plane
STP = Sole Thrust - Plane
0 1 2 3 4 km
Rocks affected by : a) the Moine Thrust
b) the Ben More Thrust
c) the Glencoul Thrust
d) the Sole Thrust
The unmoved "foreland" area
Major alkaline igneous complexes

embayment in the line of the Moine Thrust in the Assynt district is due to subsequent differential erosion and removal of various parts of the different nappe sheets. Elliott & Johnson (1980) considered the array of dislocation features in the Assynt area as part of a 'duplex' structure, i.e. "a series of curved faults asymptotically related to a higher ('roof') thrust and a lower ('floor') thrust" (Johnson & Parsons 1979, p. 11).

There has been much debate as to the age of the Moine Thrust belt and the sequence of thrust development. It seems likely that the thrust was in motion in Silurian times (van Breemen et al. 1979). Macgregor & Phemister (1972) argued that the Moine Thrust was the latest thrust and has overlapped the lower thrusts. Johnson & Parsons (1979) point out that "According to the duplex concept 'transgressions' and 'overlaps' of the lower thrust by the higher thrusts are probably illusions, the reality being that the Moine thrust evolved first, followed by the Ben More or Glencoul and finally by the Sole thrust" (p. 17).

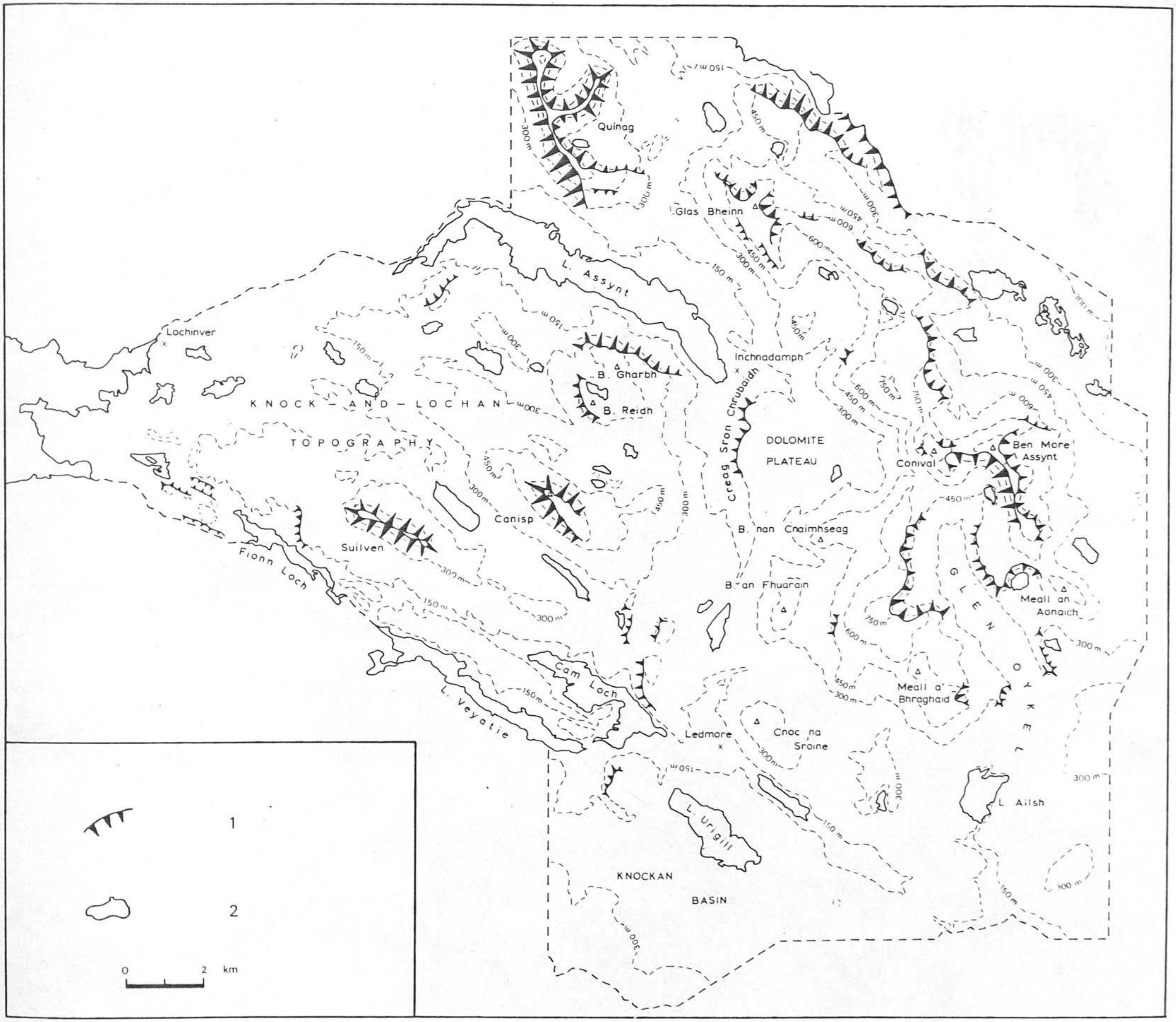
2.2 RELIEF

Although not possessing the large areas of high ground that are found farther south, the Assynt area has a varied relief that has resulted in some spectacular scenery. The region can be subdivided physiographically into a number of areas (Fig. 2.3).

In the west, glaciation of the areas corresponding to the gneissose rocks of the Lewisian complex has resulted in the distinctive scenery of numerous steep-sided knolls, with many of the intervening hollows occupied by small lochans, to which Linton (1963) gave the name 'knock-and-lochan' topography. Powerful glacial erosion and modest

Fig. 2.3 Main physiographic elements of the study area.

(diagram overleaf)


Quinag
Glas Bheinn
L. Assynt
Lochinver
Inchnadamph
B. Gharbh
B. Reidh
KNOCK-AND-LOCHAN
TOPOGRAPHY
Creag Sron Chrubaidh
DOLOMITE
PLATEAU
Conival
Ben More
Assynt
Canisp
Suilven
B. nan Cnaimhseag
B. an Fhuarain
GLEN OYKEL
Meall an
Aonaich
Fionn Loch
Cam Loch
L. Veyatie
Meall a'
Bhraghaid
Ledmore
Cnoc na
Sroine
L. Urigill
L. Ailsh
KNOCKAN
BASIN
150 m
300 m
450 m
600 m
750 m
1
2
0
2 km

fluvial erosion has resulted in a deeply dissected topography of incredible irregularity. Mostly bare rock, the relief ranges from sea level to over 350 m O.D.

Inland is a line of hills, each one orientated approximately W-E. From south to north these peaks are Suilven (731 m), Canisp (846 m), Beinn Reidh (572 m), Beinn Gharbh (540 m) and Quinag, a ridge with a number of summits ranging from 713 m to 808 m. These hills are the erosional remnants of a once continuous cover of Torridonian grits and Cambrian quartzites. All except Suilven have steep northern, western and southern sides, but their eastern sides follow the gentle incline of the Cambrian quartzite dipslope. Suilven is one of Scotland's most spectacular peaks : it is a 2 km-long ridge, steep on all sides, rising 400 m above the surrounding knock-and-lochan area. To the east of the hills is a major structural valley, stretching from Unapool and Kylesku in the north, via Inchnadamph to Ledmore in the south, a distance of 20 km. For the most part a marshy lowland area, it extends at its southern end into the Knockan basin, composed of dolomite. The eastern side of this valley, south of Inchnadamph, is a cliff, the Creag Sron Chrubaidh, which forms the western edge of a plateau of Cambrian dolomite extending 4 km to the foot of the mountain ridge to the east. With a surface at 200-250 m O.D. the plateau is capped in two places by thrust 'klippen' of Torridonian grits, forming the hills of Beinn an Fhuarain and Beinn nan Cnaimhseag. To the south of the dolomite area, the hill mass of Cnoc na Sroine (397 m) is associated with the Loch Borralan igneous intrusion.

The main mountainous area consists of a ridge trending approximately north-south, from Glas Bheinn (776 m) in the north to Meall a' Bhraghaid (688 m) in the south and Meall an Aonaich (715 m) in the south-east. The mountains approximately follow the line of 'the Assynt Culmination', the thickest area of thrust rock masses in the Moine Thrust belt (Elliott & Johnson 1980, Fig. 1), although the main peaks presently occupy an area to the east of this line because here the more resistant lithologies of the dislocated masses crop out. Hence the mountains are largely structurally controlled although glaciation and other denudation processes have shaped their present form. Craggy and steep-sided, the mountains are often plateau-like on top, except where individual peaks are connected by narrow arêtes. The ridge has a number of summits, including the highest parts of the field area in Ben More Assynt (998 m) and Conival (987 m). Upper Glen Oykel runs southwards from beneath Ben More Assynt to Loch Ailsh some 8 km away. The eastern side of the mountains for the most part comprises several large corries and associated steep rockwalls. Britain's highest waterfall, the Eas a' Chual Aluinn (NC 280277) drops 200 m over one of these abrupt topographic features, which are largely structurally controlled in the northern part of the area. The ground between the eastern side of the mountains and the Moine Thrust is characterised by irregular rocky topography or hummocky drift. To the east of the Moine Thrust, extending to Loch Shin and beyond, the topography is one of gently undulating moorland.

2.3 CLIMATE

The climate of the Assynt area is very oceanic in character. January temperatures are greater than 4° C on average, which is comparable to winter temperatures experienced in much of Wales and the West Midlands of England. The area has cool, moist summers, with a mean July temperature of less than 14° C (n.b. both values are reduced to mean sea level). Actual temperatures are much lower in localities exposed to the wind : indeed, many of the mountain tops possess actively-forming periglacial features (section 3.5).

Most of the rainfall for the area accompanies the north-westerly, westerly and south-westerly winds associated with the cyclonic weather conditions that exist for most of the year. There is a marked difference between the amount of rainfall received at the coast and that received in the mountains. The weather station at Lochinver (NC 092222, at 10 m O.D.) receives a mean annual precipitation of 1236 mm p.a., while the one just east of Ben More Assynt (NC 346194, at 370 m O.D.) receives a mean annual precipitation of 3142 mm p.a. (British Rainfall Statistics, H.M.S.O., for the period 1916-1950). There is nowhere where the winter's snow lies for long, except in sheltered locations and in gullies on the mountain slopes.

2.4 SOILS AND VEGETATION

2.4.1 Soils

Very few of the mineral soils in Scotland are derived from the *in situ* weathering of bedrock (Curtis *et al.* 1976). Most have formed on deposits that are the products of glacial or periglacial activity. Such deposits are very restricted in distribution in the Assynt area (sections 3.4 and 3.5). Shallow calcareous brown earths, based largely on fine-grained alluvial deposits, have developed in the valleys of the dolomite areas. Podzols occur on plateau areas where suitable drift deposits are available. Some very immature, skeletal soils occur on the finer-grained periglacial deposits on exposed upland areas.

In the vicinity of Inchnadamph, patches of a red soil overlying dolomite bedrock can be seen (e.g. in the road-cutting at NC 246229 and near Stronchrubie at NC 252194). The red soil (Munsell colour 2.5YR 4/4) occurs on both Grudaidh and Eilean Dubh dolomites where the more extensive drift-based podzols, with their characteristic yellowish-red B-horizons, are absent. Its red colouration is assumed to be due to the oxidation of iron in the parent material and a lack of organic carbon. In this respect the red soil is similar to 'terra rossa' soils occurring in the Mediterranean area, which have formed under conditions far warmer than those existing in Britain today. Godard (1965) suggested that these patches of red soil were vestiges of a 'terra rossa' that had formed in warmer and drier preglacial times or during an interglacial prior to the present one. Similar relict soils have been found in South Wexford, Republic of Ireland (Gardiner & Ryan 1962).

Organic soils are very extensive in the Assynt area, with ombrogenous blanket peat covering large areas of both drift deposits and solid bedrock. The presence of blanket bog "is probably a response both to climatic wetness and the impermeable glacial drifts of the region" (D.V. McVean in Burnett 1964, p. 573). The peat is often deep, with depths of up to 2.2 m being recorded on the Moine schist area east of the Oykel valley. Peat sections in the area have been examined and described by Lewis (1907) and Samuelsson (1910). Both commented on the presence of 'forest beds', containing branches and trunks of tree birches (and sometimes pine) near the base of the peat in several localities.

2.4.2 Vegetation

The prevailing vegetation of the Assynt area is blanket bog, with Trichophorum - Eriophorum species dominant at lower levels and Calluna - Eriophorum higher up the hillsides. In the wetter areas much Sphagnum moss exists, and in areas of moderate drainage Mollinia - Calluna moorland occurs. Heather becomes dwarfed by exposure at 400 - 500 m O.D., and gives way to Rhacomitrium moss heath, with Juncus trifidus and Festuca ovina at higher altitudes. Minor communities are also present, such as patchy prostrate juniper scrub on the quartzite areas, Vaccinium - Empetrum and Rhacomitrium - Empetrum heathlands, Nardus and Deschampsia cespitosa grasses and various sedges.

The outcrop of Cambrian dolomite in the Assynt area has produced a vegetation type rarely found elsewhere in the North-West Highlands. Calcareous soils and extensive grazing have led to the development of a rich sward of Agrostis and Festuca grasses, largely covered in the summer months by a thick cover of bracken. Alpine species such as Dryas octopetala occur on the dolomite area.

Although now almost completely treeless, the presence of tree stumps in the peats indicate that formerly the area was tree-covered to at least 250 - 300 m O.D. (Fig. 2.4). Some tree birches (mainly Betula pubescens spp. oderata) are sparsely distributed over the area. Most of the small islands in the lochs have a cover of tall juniper scrub, protected from grazing and fires. The land use of the area is predominantly hill-sheep farming and deer forest. Large sections of the area are currently being re-afforested by the Forestry Commission, planting lodgepole pine (Pinus oderata) and sitka spruce (Picea sitchensis).

Fig. 2.4 Photographs: (a) Suilven from Canisp;

(b) tree stumps in the peat, Glen Oykel.

(photographs overleaf)

a

b

rocks overlying the Lewisian complex, now in many places exhumed; and the west side of the Loanan valley between Ledbeg and Inchnadamph, and its northward extension, follow the Cambrian quartzite dipslope. Examples of geological control of other geomorphic features at smaller scales will be discussed subsequently.

As many of the macroscale features, and most of the mesoscale and microscale features, of the landscape have developed during the fluctuating climates of the Quaternary, discussion of the distribution of such features must allow for changes in certain of the environmental variables from those that exist at present. Analysis of the distribution of certain geomorphic features by analogy with regions possessing such features actively being formed today can give some indication of past environmental conditions in the area.

In certain cases, where the above are not the overriding controls, the distribution of particular landforms may be influenced by variables inherent in the geomorphic process that formed them. As most of the surficial features of the Assynt landscape were formed during former glacial phases, when geomorphic processes were qualitatively at their most active, it is convenient to look separately at the landforms of glacial/fluvioglacial erosion, glacial/fluvioglacial deposition and periglacial landforms, attempting to explain their distribution in terms of the above.

3.3 LANDFORMS OF GLACIAL AND FLUVIOGLACIAL EROSION

3.3.1 Corries

The Assynt area possesses several corries that fit the definition of Evans & Cox (1974, p. 151) of "a hollow, open downstream but bounded upstream by the crest of a steep slope (headwall), which is arcuate in plan around a more gently sloping floor". All are found in the mountainous area stretching from Quinag to Meall a' Bhraghaid (Fig. 3.1). To these classic corrie forms can be added the long sections of steep rockwalls on the east side of Beinn Uidhe and Beinn an Fhurain which have also acted as glacier source areas.

Except for the shallow corrie to the SE of Glas Bheinn, the corries show a certain amount of lithological control in that all are floored by Lewisian gneiss, whereas their headwalls are either quartzite or Torridonian sandstone. Corries to the east of Ben More Assynt are excavated entirely in one lithology, although often one or more of their sides has formed along a line of weakness, such as a fault or a dyke. This accords with the work of Thompson (1950), who concluded that the detailed development of thirteen corries farther north was closely related to the distribution of different rock types and their relative resistances to erosion.

In Scotland as a whole, Sissons (1967, Fig. 21) found that corrie orientations between NNW and ESE were favoured. In the Assynt area, sixteen out of twenty corries and rockwalls have aspects between N and SE. Corries in the north of the study area face from just west of north to NE, whereas in the south of the area aspects between E and SE are favoured (Fig. 3.1). Climatic controls such as those discussed by Sissons (1967, pp. 57-63) doubtless influenced the distribution and aspect of corries in Assynt, but with such a small sample in a restricted area it is not possible to show any meaningful statistical relationships.

3.3.2 Glacial troughs and rock basins

Glen Oykel is the best example of a glaciated valley in

Fig. 3.1 Landforms of glacial and fluvioglacial erosion in the Assynt area.

Key: 1. Corries and rockwalls.

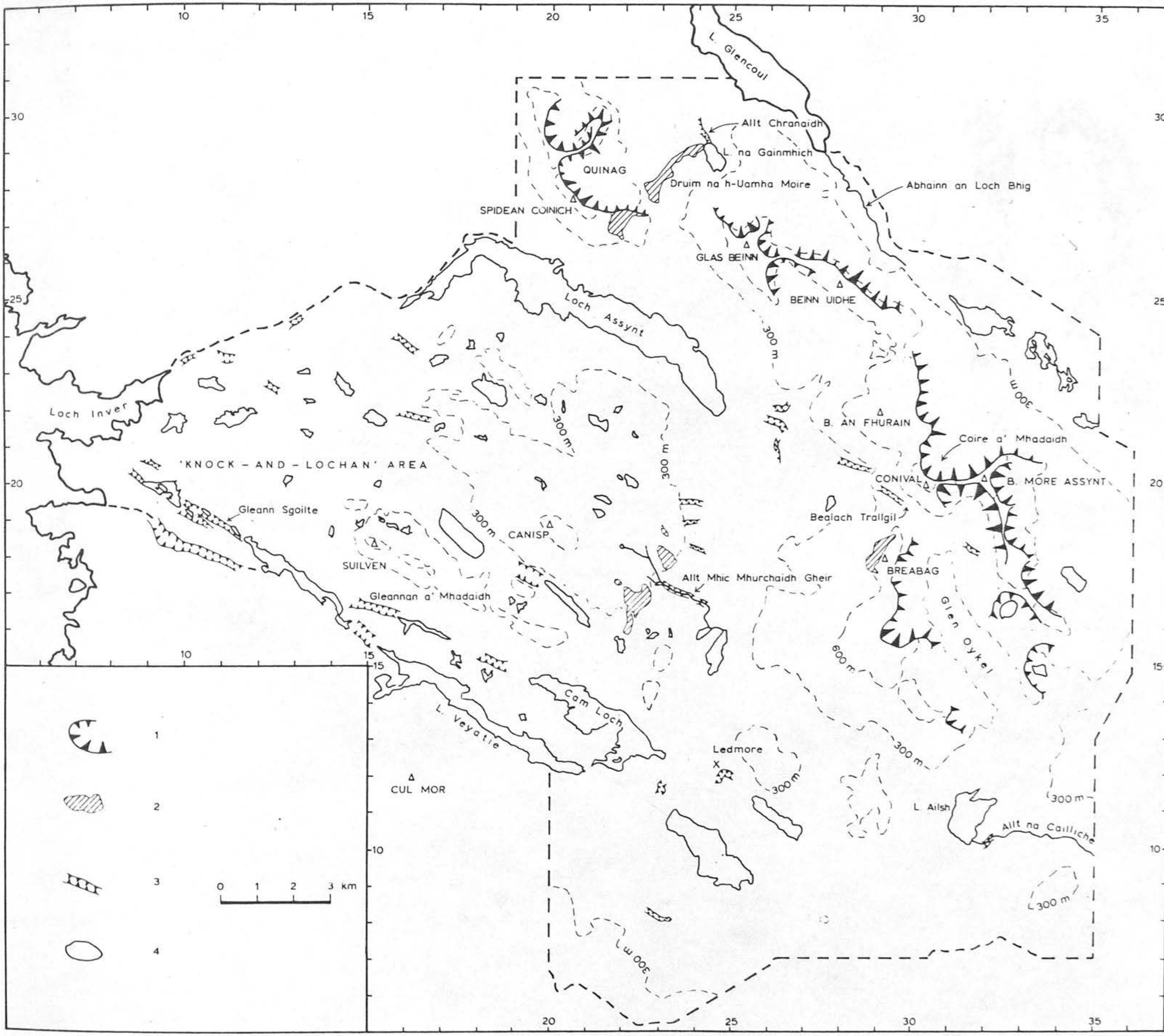
2. Striated pavements.

3. Meltwater gorges.

4. Lochs.

Contours at 300 m intervals.

(diagram overleaf)



L. Glencoul
Allt Chranaidh
L. na Gainmhich
QUINAG
Druim na h-Uamha Moire
SPIDEAN COINICH
Abhainn an Loch Bhig
GLAS BEINN
BEINN UIDHE
Loch Assynt
300 m
Loch Inver
B. AN FHURAIN
Coire a' Mhadaidh
'KNOCK - AND - LOCHAN' AREA
CONIVAL
B. MORE ASSYNT
Gleann Sgoilte
Bealach Trallgil
CANISP
SUILVEN
BREABAG
Allt Mhic Mhurchaidh Gheir
Gleannan a' Mhadaidh
Glen Oykel
600 m
Cam Loch
L. Veyatie
Ledmore
CUL MOR
L. Ailsh
Allt na Cailliche
1
2
3
4
0 1 2 3 km

the study area, exhibiting classic glaciated transverse and longitudinal profiles. Deep incision is also shown by the valley of the Abhainn an Loch Bhig and its extension into the fjord of Loch Glencoul. Glen Oykel seems to show no geological control, but the Abhainn an Loch Bhig follows the line of a major thrust plane.

Evidence for glacial erosion in the area is also provided by the large number of rock basins, often completely enclosed but sometimes interconnected by streams. The lakes that occupy most of these rock basins vary in size from the major lochs (e.g. Loch Assynt, Cam Loch and Loch Veyatie) to the numerous small lochans that may be seasonally dry (Fig. 3.1).

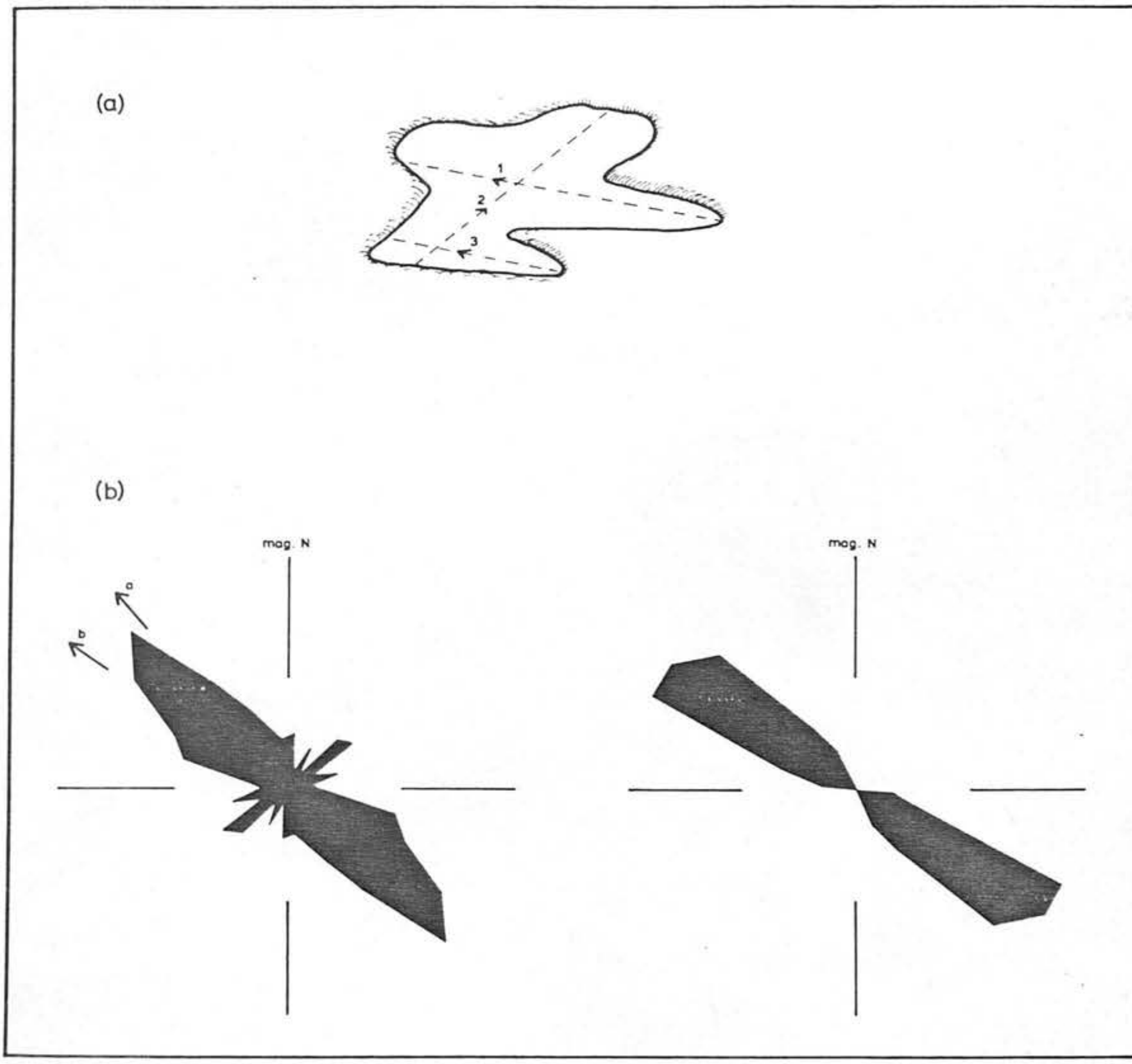
Since maps indicate an apparent alignment of lochs, especially in the knock-and-lochan area, an attempt was made to quantify this pattern using the O.S. 1:25,000 (Second Series) maps. Only those lochs and lochans on the Lewisian rocks of the unmoved 'foreland' area (Fig. 2.2) whose A:B axes have a ratio of greater than, or equal to, 3:2 were considered, the orientation of the longest axis being measured. For certain irregularly-shaped lochs it was possible to define more than one major axis, in which case the orientations of the two or more major axes were recorded (e.g. Fig. 3.2(a)). A total of 259 axis orientations was obtained for 226 lochs. In order to assess the effects of geological weaknesses on the orientation pattern, the orientation of all the basic and ultrabasic intrusions shown crossing the same Lewisian rocks on the Assynt district geological map were measured; this gave 298 values. Fig. 3.2(b) shows the results of these analyses in the form of rose diagrams and compares them to the orientation of the long axes of closed contours on Suilven and Canisp. A strong orientation of the lochs is apparent, with a major mode trending SE-NW and a small secondary mode trending NE-SW. The major

Fig. 3.2 Quantification of the orientation of rock-basin lakes in the Assynt area.

(a) The measured axes (1, 2 and 3) in the case of an irregularly-shaped loch.

(b) (Left) Rose diagram showing the orientation of lochs and closed contours on Canisp (a) and Suilven (b); (Right) Rose diagram showing the orientation of igneous intrusions.

(diagram overleaf)


(a)
1
2
3
(b)
mag. N
mag. N

mode is very similiar to that obtained for the dykes and the secondary mode may relate to the pattern of fault lines that crosses the area. It is inferred that the orientation of Suilven and Canisp, aligned orthogonal to the main geological boundaries, reflects the direction of former ice movement in the area: their orientation is very similar to that of the lochs. It therefore seems likely that the strong orientation of rock basins in the knock-and-lochan area is due to the exploitation of lines of weakness in the local gneissose rocks by glacier ice which tended, in this region, to flow in a similar direction. The geological influence is considered to be the more important of the two factors, as evidence from glacial erratics and striae (chapters 4 and 5) suggests that the most recent direction of ice flow over the knock-and-lochan area followed a more westerly course than the orientation of Suilven and Canisp. Many workers in other areas (e.g. Auden 1954; Sissons 1967; Niini 1968) have commented on the relation of glacial troughs and rock basins to lines of geological weakness.

3.3.3 Ice-moulded bedrock

Ice-moulded bedrock landforms, of which roches moutonnées are the most often quoted, are present in many parts of the study area. Most standard textbooks state that the orientation of the long axis of a roche moutonnée, together with a consideration of the smoothed proximal ('stoss') side and 'plucked' distal ('lee') side, indicates the direction of former ice movement in an area. It is suggested here that such stoss-and-lee features are not unequivocal directional indicators of former ice flow, for local geological structure has to be taken into consideration.

In attempting to study the direction of former ice flow over the knock-and-lochan area, where striae are almost absent, the author considered using the orientation of roche moutonnée forms. However, it became apparent that such features are strongly influenced by the structure of the Lewisian rocks. Although in this area most 'plucked' faces point westwards, indicating a westward ice movement later supported by a study of glacial erratics (chapter 4), the long axes of many ice-moulded forms clearly follow igneous intrusions, fault lines and joint patterns. Also, in instances where the long axis of the landform is not clearly defined, there is a tendency for the field-worker's eye to pick out an axis that fits other alignments in the topography, which are often structurally controlled as has already been shown. In cases where relief elements possess bedding planes which dip in a down-ice direction, glacially-freshened surfaces facing up-ice can be produced; such surfaces often occur around crag-and-tail features (e.g. Arthur's Seat and Salisbury Crags in Edinburgh), but in some instances can produce ice-moulded bedrock forms which are the exact reverse of the classic roche moutonnée. This has been independently shown in recent work by Gordon (1981). Therefore, without a consideration of local bedrock structure, it is not possible to say with any degree of certainty which direction the ice moved across a region in the past. It has also been noted that subglacial changes in effective normal stress can produce both normal and reversed roche moutonnée forms (e.g. Derbyshire et al. 1979, p. 239). In view of these difficulties, it was decided that other evidence would be used to determine detailed ice-flow patterns across the study area.

Strong grooving of Cambrian quartzite surfaces occurs between Cam Loch and Loch Veyatie. A series of ridges, trending ESE-

WNW, with intervening peat-filled troughs, is closely paralleled by the striae on the ridge surfaces, indicating a subglacial origin. Similar grooving of quartzite can also be seen on the eastern slopes of Cul Mor, to the SW of the study area.

3.3.4 Striated Surfaces

As the area possesses many fine striated rock surfaces, glacial striae were studied in detail (chapter 5). However, a brief mention must be made of the distribution of several areas of continuous smooth rock surfaces showing much striation, hereafter referred to as 'striated pavements' (cf. limestone pavements). Such striated pavements are well represented between 425 and 500 m on the eastern slope of Spidean Coinich on the Quinag ridge, 275 and 320 m on Druim na h-Uamha Moire, 580 and 640 m on Breabag and 300 and 450 m on the eastern slopes of Canisp (Fig. 3.1). All these striated pavements are composed of quartzite, which in all cases dips gently up-ice, presenting a smooth slope up which the ice flowed. The distribution of these striated pavements is altitudinally controlled, lying between the local upper limit of blanket peat and the local lower limit of periglacially frost-shattered bedrock.

3.3.5 Meltwater gorges

Fig. 3.1 shows the main examples of deep, narrow gorges in the study area. Those in the knock-and-lochan area have been excavated along igneous intrusions, faults and shatter-belts running through the Lewisian rocks. Many of these deep channels are floored with peat or carry only small streams. Several are blocked by talus that has fallen from the channel sides (e.g. Gleann Sgoilte at NC 109190). The deep channels running down the dip-slope of the quartzites on the eastern side of Canisp are all occupied by relatively undersized streams, except for a few tributary channels which are normally dry. None of this group of gorges seems to be influenced by the geological structure; the upper reaches of the Allt Mhic Mhurchaidh Gheir follow a fault line, but here the stream course is not particularly incised.

In the northern part of the field area, the Allt Chranaidh occupies a deep gorge after leaving Loch na Gainmhich; this gorge follows a fault line. The Allt a' Bhealaich is deeply incised into quartzite after leaving the Bealach Trallgil beneath Conival. The deep channel and the Bealach Trallgil follow the same fault line, the continuation of which can be traced across the head of the Oykel valley to another marked incision carrying a tributary of the Oykel. The Allt na Cailliche occupies a deep cutting in the Moine schist shortly before it enters the southern end of Loch Ailsh. Finally, the Ledmore River flows through a deep meandering channel near Ledmore Cottages. Other deep channels, often dry, occur on the dolomitic limestone areas; these are considered in chapter 9.

It is difficult to reconcile the size of many of the channels mentioned above (e.g. up to 15 m deep in the case of the Allt Mhic Mhurchaidh Gheir and 45 m deep for Gleannan a' Mhadaidh, south of Suilven) with the size of the streams occupying them. It is therefore likely that many of the gorges were initiated by glacial meltwaters. Much of the incision, which clearly favoured structural weaknesses in the rocks as the deeper gorges follow such features, may have been achieved subglacially, but larger, more powerful proglacial rivers carrying debris-rich meltwaters on deglaciation probably further excavated some of the features.

3.4 LANDFORMS OF GLACIAL DEPOSITION

Glacial deposits in the Assynt area can be subdivided into two categories : till, and erratic blocks on bare rock surfaces. The latter are considered in chapter 4; hence this section is concerned with the characteristics of the till, its distribution and the landforms composed of it.

3.4.1 Characteristics of the till

Till in the Assynt area is generally very blocky in character, with little matrix (cf. the granular or 'clast dominated' till of Derbyshire et al.(1976)). This reflects the major lithologies, which are hard crystalline rocks, more susceptible to quarrying processes than subglacial abrasion. The tendency for quartzite to break into joint-controlled blocks, and its durability, means that quartzite tends to dominate the lithological composition of the till, regardless of the local rock type. The till has normal characteristics such as striated clasts generally of subangular and subrounded shape. Some analyses of till from the dolomitic limestone area are discussed in chapters 9 and 10.

3.4.2 Distribution of till in the Assynt area

Over much of the field area glacial deposits are absent or thin; where present they occupy small hollows between areas of upstanding bedrock. In certain localities, however, thick glacial deposits occur and are associated with distinctive landforms.

(a) Lateral and terminal moraines (Fig. 3.3). Well developed end moraines occur in the Loanan valley opposite the Creag Sron Chrubaidh

Fig. 3.3 Landforms of glacial and fluvioglacial deposition in the Assynt area.

Key: 1. Lateral and terminal moraines.

2. Areas of hummocky moraine.

3. Fluted moraines.

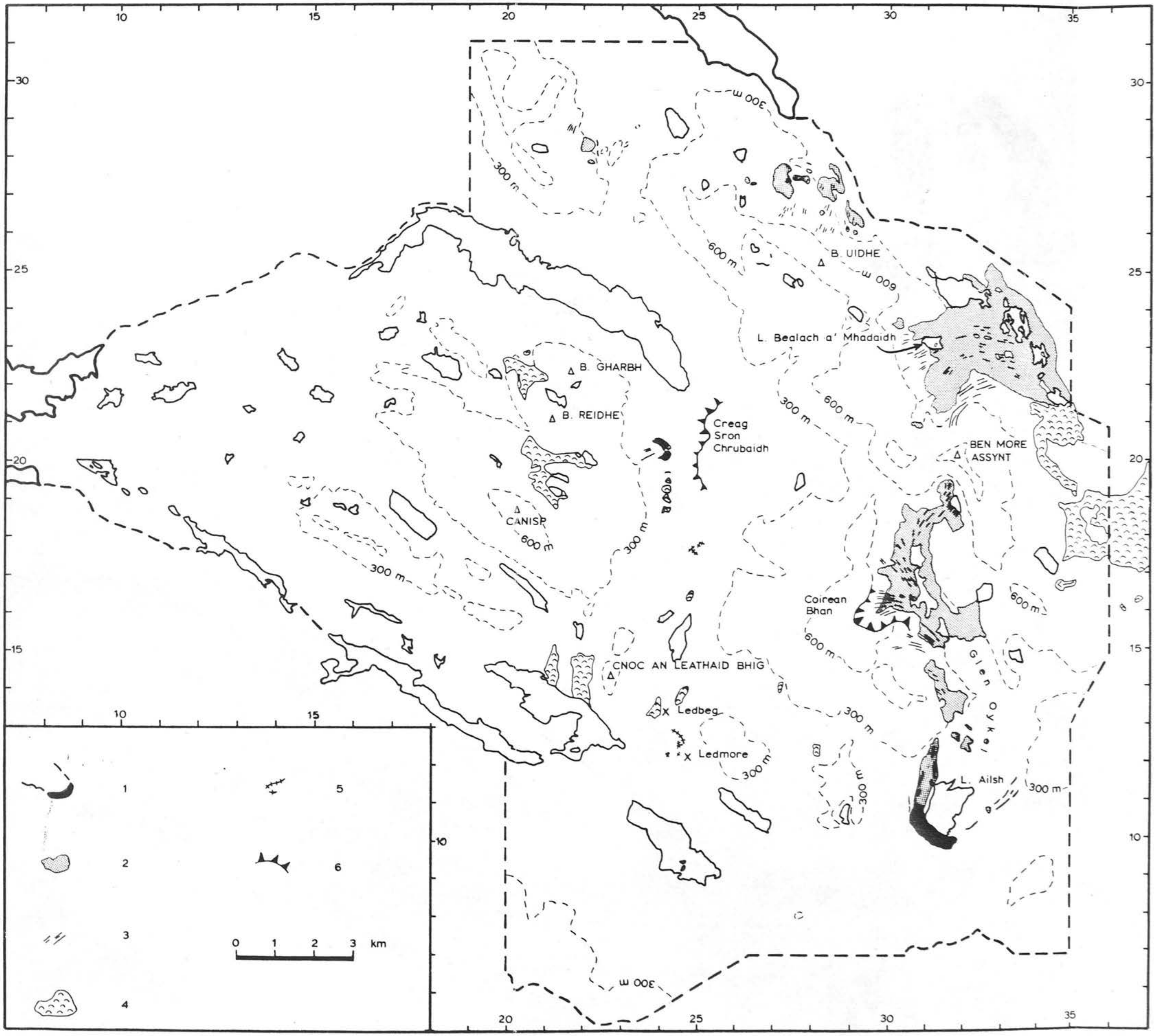
4. Till mounds.

5. Esker-like ridges.

6. Crags mentioned in text.

Contours at 300 m intervals.

(diagram overleaf)



B. UIDHE
L. Bealach a' Mhadaidh
B. GHARBH
B. REIDHE
Creag Sron Chrubaidh
BEN MORE ASSYNT
CANISP
Coirean Bhan
CNOC AN LEATHAID BHIG
Ledbeg
Ledmore
Glen Oykel
L. Ailsh
300 m
600 m
1
2
3
4
5
6
0 1 2 3 km

and across the southern end of Loch Ailsh. At the former locality there are three closely-spaced ridges, the steep distal side of one being 5 m high. Long lateral moraines can be traced up-valley from the Loch Ailsh end moraine, with roughly parallel ridges recording successive stages in the retreat of a glacier that occupied Glen Oykel. Other mounds and short ridges interpreted as lateral and terminal features are detailed in section 6.3.

(b) Hummocky and fluted moraines. Distinctive areas of vegetated, irregular morainic hummocks are shown in Fig. 3.3. In the literature the differentiation of hummocky moraine areas from those of other morainic mounds has always been a qualitative one, based on the 'freshness' of appearance of the former (Sissons 1967, p. 139; 1974). Lack of a definition of the term has led to controversy (e.g. Sugden 1970, 1973; Sissons 1973). Nevertheless, a distinction can be made between the areas labelled 'hummocky moraines' and areas of 'till mounds' in Fig. 3.3.

The hummocky moraines are often conical in shape, although ridges also occur; deep depressions are usually found amongst them. In many instances large boulders litter their surfaces. The usual interpretation of these features is that they are formed by supraglacial deposition on rapid ice decay (e.g. Boulton 1972), but this may have to be revised in the light of recent research which indicates that at least some hummocky moraines are deposited subglacially (D.M. Hodgson, pers. comm.). Large hummocky moraine areas occur in Glen Oykel and to the north of Ben More Assynt (Fig. 3.3), with smaller areas to the north of Beinn Uidhe and in the corrie NE of Spidean Coinich.

Both within, and in close proximity to, the areas of hummocky moraine are linear or curvilinear ridges. These fluted moraines are present in great numbers in some localities (Fig. 3.3). They vary in height from a few tens of centimetres to several metres and are up to 700 m long. They are not always clearly discernible in the field, but are clear on aerial photographs from which they can easily be mapped. Formed subglacially by active ice, the fluted moraines are aligned parallel with the direction of glacier flow. They show a variety of forms, from a thin smearing onto underlying glacial deposits (e.g. those emanating from Coirean Ban at approx. NC 300163), short 'humpbacked' forms (e.g. those south of Dubh Loch Mor beneath Ben More Assynt) and long steep-sided ridges (e.g. at approx. NC 305168 in the Oykel valley).

(c) Miscellaneous areas of mounds. Mounds of till that have a more subdued and rounded form than hummocky moraines are present at several localities (Fig. 3.3). They occur west of Beinn Gharbh, between Beinn Reidhe and Canisp, west of Cnoc an Leathaid Bhig, along the Loanan valley, at several locations in the valley of the Ledbeg River and they occupy a large area between the Assynt mountains and the northern end of the Cassley valley.

Other areas contain mounds and ridges which may have a fluvioglacial origin. Fig. 3.3 shows some of the sinuous ridges which have the form of eskers. Of particular note are those in the vicinity of Ledmore, where a small section (at NC 24231238) cut through one of these features by the A 837 road shows that it is composed of clean, rounded gravel.

3.4.3 Factors influencing the distribution of till and moraines

The most important factors influencing the distribution of till landforms in the study area are the local geology and the distribution of Lateglacial glaciers (section 6.3).

The predominance of quartzite clasts in the local till has already been mentioned. The presence of such a relatively durable material gives the Assynt till its characteristic blocky texture. There is a striking similarity between the distribution of quartzite bedrock (Fig. 4.4) and the extent of hummocky moraines in the Oykel valley and north of Ben More Assynt. It is suggested that the incorporation of large amounts of quartzite clasts into the till in these areas gave the till a large degree of frictional strength which allowed the hummocks to form and largely to retain their shape after deglaciation. However, hummocky moraines to the north of Beinn Uidhe occupy an area of Lewisian bedrock, as do those at the very head of Glen Oykel and some to the north of Ben More Assynt (around Loch Bealach a' Mhadaidh). Although these areas of hummocky moraines include some local material, they are still largely composed of clasts from quartzite outcrops up-ice or on high ground surrounding the glacial source areas.

It is possible to extend this hypothetical explanation of the distribution of hummocky moraines to include the other areas of till mounds mentioned in section 3.4.2 (c). Although these areas are largely found on lithologies that are unlikely to produce a blocky till, and hence mounds, due to their argillaceous nature or low durability (e.g. the pelitic schists and dolomites), they are all composed of such a till, rich in quartzite with a sandy matrix, which originated from the well-jointed and more granular lithologies up-ice.

The fresher appearance of hummocky moraines (cf. morainic mounds) is most probably a function of age, with hummocky moraine areas relating to a more recent glacial phase, during which glacier ice did not extend onto the areas occupied by the till mounds.

A further point of interest is the presence of large numbers of Lewisian gneiss blocks associated with some of the hummocky moraine areas. On the knock-and-lochan area, composed chiefly of gneiss, such blocks are absent except where they have fallen as talus from local free faces. This suggests that widespread glacier ice, as under mature ice sheet conditions, is incapable of removing Lewisian gneiss blocks by some sort of 'plucking' process, whereas local corrie and valley glaciers, such as might exist in an early phase of ice-sheet glaciation, crossing gneiss bedrock freshly weathered and loosened in a previous interstadial or interglacial, are able to incorporate such material.

3.5 PERIGLACIAL LANDFORMS

This section discusses the distribution of both fossil periglacial landforms relating to severe periglacial conditions that existed in the past and those that owe their existence to the harsh conditions occurring on the higher ground at present (Fig. 3.4).

3.5.1 Types and distribution of periglacial landforms in the Assynt area

(a) Landforms relating to frost-shattering processes. These form by far the largest group of features, including blockfields, blockslopes and roughened bedrock surfaces.

Frost-shattered debris mantles most ground above 480 m O.D. and occasionally reaches down to *c*. 330 m (Fig. 3.4). The most

Fig. 3.4 Distribution of certain periglacial and slope features in the Assynt area.

Key: 1. Free faces.

2. Bedrock outcrops roughened by frost-shattering.

3. Blockfields.

4. Vegetated blockfields.

5. Active talus slopes.

6. Debris slopes.

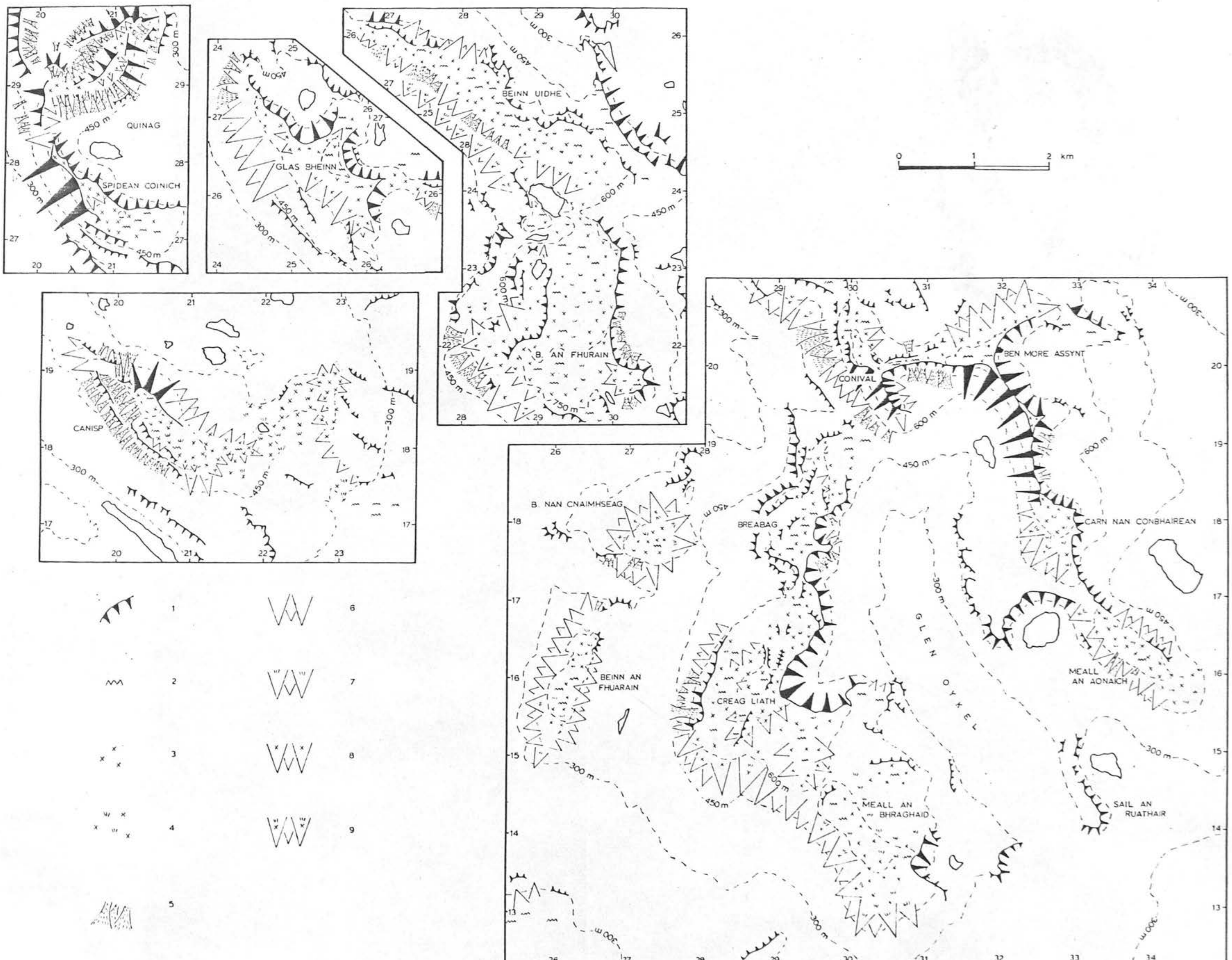
7. Vegetated debris slopes.

8. Blockslopes.

9. Vegetated blockslopes.

Contours at 150 m intervals.

(diagram overleaf)


QUINAG
SPIDEAN COINICH
GLAS BHEINN
BEINN UIDHE
B. AN FHURAIN
CANISP
CONIVAL
BEN MORE ASSYNT
B. NAN CNAIMHSEAG
BREABAG
CARN NAN CONBHAIREAN
BEINN AN FHUARAIN
CREAG LIATH
GLEN OYKEL
MEALL AN AONAICH
MEALL AN BHRAGHAID
SAIL AN RUATHAIR
0 1 2 km

impressive of these inactive blockfields occur on quartzite outcrops (Fig. 3.5(a)) where the heavily-jointed bedrock is in places covered with angular blocks 1-2 m across to depths of at least 3 m. The same debris forms quite steep blockslopes in many places, the large interlocking blocks allowing a high angle of rest. Frost - shattering of more granular rock types, such as the quartz-syenites of the Loch Ailsh and Loch Borralan intrusions, and the Torridonian sandstones and grits, tends to produce a thinner debris mantle of smaller, less angular particles.

Bedrock surfaces at all levels exhibit the effects of frost-shattering processes, such as the widening of joints, cleavage and bedding planes and the removal of crack-bounded blocks. Many of the harder lithologies, such as quartzite, still retain evidence of previous glaciation (ice-moulding, striations, etc.) in areas where complete shattering has not been accomplished. Postglacial granular disintegration by both frost and chemical weathering processes has tended to remove all such surface features from less resistant lithologies such as the Torridonian sandstone.

(b) Landforms of periglacial mass movement processes. Included here are all geomorphic features in the study area that can be attributed to gelifluction and frost creep. It is appreciated that other slope processes which are not strictly due to periglacial conditions may well have affected such landforms, but they are included here for ease of description.

Beneath all free faces in the study area, debris slopes have accumulated. Particle sizes range from coarse sand up to blocks a metre across. Many such debris slopes are still actively accumulating, but a large number of them have been stabilised by vegetation and are

Fig. 3.5 Photographs: (a) blockfield between Creag Liath and Meall Diamhain;

(b) frost-shattered erratic of Canisp porphyry on the eastern slopes of Canisp.

(photographs overleaf)

a

b

hence fossil.

Many varieties of terrace and lobe form occur on the steeper slopes where there is a cover of weathered debris. Above 640 m on Carn nan Conbhairean, and on the NW slope of Meall an Aonaich, terraces with small risers (10-15 cm high) run obliquely down the local slope. They are composed mainly of gravel-sized particles with occasional angular clasts up to 30 cm in diameter. These features have only strips of vegetation along the top of each riser and are therefore probably still active. Above 790 m, on Carn nan Conbhairean, boulders form garlands of lobate features with risers of approximately 30-50 cm. Many of the boulders have their long axis pointing down the slope, those of elongate shape often dipping steeply in a downslope direction. These features are probably fossil gelifluncted boulder lobes, relating to former severe periglacial conditions. Similar forms are present in areas covered by quartzite blockfields, but they are less conspicuous due to the completeness of the cover of weathered debris (e.g. at NC 285164 on Creag Liath).

Near the summit cairn of Canisp, at about 820 m O.D., well-formed gelifluction lobes occur. They are composed of mainly cobble- and gravel-sized clasts which are often highly rounded due to microgelivation processes. Their downslope ends are turf-banked. These lobes are 100-150 cm across at their widest point, and are therefore smaller in size than the boulder lobes on Carn nan Conbhairean and on Canisp itself. Two examples of ploughing blocks were found at NC 20451842, on a slope of about 30°, with furrows about 10 m long. Terrace forms include both turf-banked terraces running parallel with the contours (at NC 208180) and oblique interconnecting types (at NC 207183).

(c) Patterned ground. Patterned-ground phenomena in the study area can be divided into two types : those comprising angular blocks of cobble-size or greater, which are often partly vegetated and are therefore considered to be inactive, and those of smaller particle sizes which show a greater degree of sorting and are probably actively forming.

All examples of the former, inactive, type occur amongst the quartzite blockfields on the main mountain ridge, where a suitably large amount of coarse weathered debris exists. Coarse boulder polygons, generally showing a minimum of sorting, with diameters of 100-350 cm, give way to elongated forms and coarse stone stripes as the angle of slope increases away from the flattish mountain tops. Good examples exist at 760 m on Glas Bheinn (near the summit cairn), over 670 m on Beinn Uidhe (e.g. NC 278256 and NC 284251), over 730 m on Beinn an Fhurain (e.g. NC 288217 and NC 298221) and at 670 m on Creag Liath (NC 289167). Boulder polygons were found at only 590 m on Canisp (NC 212181) and 620 m on Meall a' Bhraghaid (NC 303137).

Small circles and polygons occur in many places where frost-shattered debris is less coarse and has a fine matrix. Such situations often occur where felsite intrusions cross the shattered blockfield areas. These small active polygons are commonly up to 50 cm across. In some cases, mosses have colonised the finer central parts of the polygons giving a 'polka dot' appearance to the local vegetation. Examples occur at 600 m in the col between Glas Bheinn and Beinn Uidhe, at 620 m on Meall a' Bhraghaid (NC 302138), at 700 m on Meall an Aonaich (NC 337163) and above 760 m on Beinn an Fhurain (e.g. NC 291215).

(d) Aeolian features. Wind blowing across certain of the more vegetated summits and through cols between them has selectively stripped the vegetation cover to form deflation stripes, where the underlying regolith shows through (e.g. at NC 308137). Almost completely bare deflation surfaces occur on some of the Torridonian sandstone hills. Areas of blown sand exist near the summit cairn of Glas Bheinn and on Quinag; similar sand sheets occur on Cul Mor and An Teallach, to the south of the study area.

3.5.2 The significance of periglacial features in the Assynt area

It has been shown by recent work (e.g. Ballantyne 1981; Chattopadhyay 1982) that some periglacial landforms are actively forming at present on many Scottish mountains. These active periglacial landforms include small stone polygons, ploughing blocks, solifluction terraces and some of the smaller solifluction lobes (Sissons 1976, p. 108). Larger forms, such as boulder lobes, the larger solifluction features, boulder polygons and blockfields, appear to be relict phenomena.

Several workers have suggested that some of these fossil features may have been active during the Postglacial period. Sugden (1971) obtained radiocarbon dates of 4,880 $\pm$ 135 and 2,680 $\pm$ 120 years B.P. for organic material buried by solifluction lobes in the Cairngorms. A date of 5,145 $\pm$ 135 years B.P. was obtained from material in a similar position on Arkle in Sutherland (White & Mottershead 1972). King (1971, 1972) suggested that stone-banked solifluction lobes and boulder polygons had been active in the Little Ice Age of the eighteenth and nineteenth centuries A.D., based on lichenometric studies. In the Assynt area, frost-shattering of glacial erratics (Fig. 3.5(b)) and glaciated pavements attributed to the last ice sheet (chapter 6) indicates that periglacial conditions have existed since then. It has been suggested (Sissons 1976, pp. 112-3; Gray & Lowe 1977; Sissons 1979b) that many of these fossil periglacial features date from the Loch Lomond Stadial (c. 11,000-10,000 years B.P.) as they complement, yet are not found within, the areas covered by glacier ice of the Loch Lomond Advance.

It is unlikely that Late Pleistocene periglacial environments in middle latitudes were comparable to those that exist in high latitudes today, because of variations in the solar radiation pattern and amount of insolation, and differences in the atmospheric circulation due to the existence of the large ice sheets (summarised by French 1976, chapter 11). Therefore, inferences about environmental conditions at the time these relict features were active, by analogy with similar features in modern periglacial areas, can only be tentative. None of the fossil periglacial features described above from the Assynt area is unequivocally indicative of permafrost conditions (French 1976, p. 236), although all indicate intense freeze-thaw activity on a scale much greater than that occurring today. However, evidence of the former existence of permafrost is provided by ice-wedge pseudomorphs found elsewhere in Scotland in contexts associated with both the last ice sheet and the Loch Lomond Advance (Sissons 1974; 1976, pp. 110-111), implying mean annual temperatures no higher than -1°C.

CHAPTER 4: THE DIRECTION OF FORMER ICE MOVEMENT : a) GLACIAL ERRATICS

4.1 INTRODUCTION

Glacial erratics are rock fragments that have been transported from their source onto an area of different lithology by the action of glaciers. The recognition that the erratic blocks widely distributed over areas outside the limits of present day glaciation were in fact evidence of former glaciation was an important factor in the general acceptance and development of the 'Glacial Theory' proposed by Venetz, de Charpentier and Agassiz in the first half of the nineteenth century (Price 1973, p. 3).

The voluminous literature on glacial erratics has recently been reviewed by Shakesby (1977); hence only an outline of the main conclusions from past work on this topic is given here.

4.1.1 The origin of glacial erratics

As well as the contribution of fresh bedrock fragments, erratics can be derived from older tills and other pre-existing deposits, such as beaches and stream beds. Nevertheless, freshly-eroded bedrock is probably the most important source of erratic material, being 'plucked' and entrained subglacially by various regelation processes, though a complete understanding of how this is achieved has been hampered, as direct observation of these processes has been minimal. Material falling onto the surface of a glacier from surrounding high ground can be transported supraglacially or englacially, such material following a variety of flow paths through the ice (Boulton 1978). There has been much debate as to how much preparation of the bedrock has to occur prior to glaciation before glacier ice can remove blocks of material, and how much loosening of fresh bedrock can be achieved by the ice itself (summarised by Shakesby 1977, pp. 14-17, and Embleton & Thornes 1979, p. 283).

Once entrained, a small proportion of erratic fragments may be carried exceptionally long distances. R.P. Goldthwait (in Goldthwait (ed.) 1971, p. 9) mentioned that certain clasts in tills in Ohio, Indiana and Illinois have travelled 480-800 km. In the British Isles, microgranite erratics from Ailsa Craig, in the Firth of Clyde, have been found in till in South Wales and southern Ireland, having travelled at least 500 km (Sissons 1976, p. 73)

4.1.2 Distribution of glacial erratics

Glacial erratics are widespread in formerly glaciated areas. Their distribution is not restricted to lower ground for they have often been noted on hills and mountains down-ice from, yet high above, the outcrops from which they originated. Several remarkable examples of the uplift of Scottish erratics above their source areas are given by Sissons (1967, p. 83), including Loch Doon erratics on Merrick (150 m of uplift in about 1 km), Lennoxtown essexite erratics on parts of the Kilsyth Hills (over 215 m of uplift in 3.5 km), and Rannoch Moor granite erratics on Schiehallion (over 600 m of uplift in 20 km). Shakesby (1977, pp. 18-20) has summarised the various hypotheses seeking to explain such uplift of material by glacier ice. Most studies have been based on observations on valley glaciers and relate to some form of shearing in the ice bringing up subglacial material to the surface of the glacier, usually under conditions of compressive flow. However, all the high-level erratics referred to above were carried to their present positions by former ice sheets.

In situations where ice has been able to ascend fairly gentle slopes, the distribution of high-level erratics can be explained by sub- or en-glacial transportation, but problems of interpretation occur when erratics appear to have been moved up steep slopes or even free faces (as in the present study), so that a large amount of uplift has occurred over a relatively short horizontal distance.

Another problem relevant to interpreting the distribution of erratics is that their present positions represent the culmination of an infinite number of changes in ice direction, during the various phases of the last or any number of previous glaciations. Notwithstanding these largely indeterminable factors, the final distribution of erratics in an area is highly dependent on the ability of the various rock fragments to survive erosion during transport, and any subsequent post-depositional weathering processes. The distribution of erratics is further restricted by the maximal extent of the former ice mass that transported them.

The variables affecting the dispersal of erratics over an area, by glacial processes, can be listed as the following:

i) geological controls, e.g. degree of jointing and faulting of the bedrock, its petrology and susceptibility to glacial erosion,

ii) the spatial distribution of the various lithologies, and the size of their outcrops,

iii) the distribution of older unconsolidated material,

iv) glaciological controls, such as ice thickness, velocity and thermal regime,

v) the direction of ice flow with respect to the underlying topography,

vi) the transportational flow paths taken in the ice by the erratics, and the distance actually travelled, both of which affect the amount of potential erosion during transportation.

If an outcrop is of limited size, composed of a hard rock type, and distinctive enough for fragments of it to be recognisable in the field, a detailed survey can result in a map of the down-ice distribution of erratic material from its source. The form that such a distribution usually takes is known variously as a boulder train, an indicator train or an erratics train. Such erratics trains are characteristically fan-shaped, with the apex of the fan centred on the source area. There is much variation in the angle of fanning-out of erratics trains, dependent on the degree of divergence of ice flow, the amount of redistribution by meltwaters on deglaciation, or the variations in direction of ice movement due to shifting centres of ice dispersal and the varying influence of the underlying topography. Hence erratics trains can fan out widely, such as the one studied in the Waterloo area of Wisconsin by Buell (1895), or they can be almost linear, as in the case of the Snake Butte boulder train (Knechtel 1942). Gillberg (1965), working in southern Sweden, described 'finger-shaped' dispersal patterns following the main valleys. Shilts (1973a) discussed similar patterns occurring in the Lac-Megantic area of Canada, though in this case the erratics trains seemed to narrow in a down-ice direction.

It has long been recognised that erratics generally decrease in abundance away from the source outcrop and away from the main axis of the erratics train. Krumbein (1937) was the first to attempt a mathematical definition of this phenomenon, using a negative exponential function. Subsequent workers fitted Krumbein's curve to the distribution of erratic material in tills, attempting to explain variations in terms of glaciological and geological variables (e.g. Dreimanis 1956; Gillberg 1965, 1967a, 1967b; Dionne 1973b).

The uses to which studies of glacial erratics have been put is basically threefold:

i) as an aid to the geological mapping of bedrock in areas covered by thick glacially-derived deposits (e.g. Mutanen 1971),

ii) as a technique in prospecting for various ores and minerals of economic value in sparsely populated and relatively unexplored areas (e.g. Sauramo 1924; Grip 1953; Dreimanis 1958; Shilts 1971, 1976, 1977; Nichol & Bjorklund 1973),

iii) as a means of determining the direction of ice flow and the character of glacial dispersal patterns in formerly glaciated regions (e.g. Dreimanis 1956; Okko & Peltola 1958; Shilts 1973b; Minell 1978).

The present study falls into the third of the above categories.

4.1.3 The suitability of the Assynt area for a study of glacial erratics

Before discussing the methodology and the results of the study of glacial erratics in Assynt, it is necessary to outline to what extent the area is suitable for such a study, and the basic aims of the work.

The geology of Assynt has been mapped in detail (e.g. Peach et al. 1907). The area possesses some distinct lithologies which have provided easily recognisable erratics. Since the early work of the Geological Survey in the area the general direction of ice movement has been known and it has been suggested that the former ice divide lay across part of the study area (Peach & Horne 1892a). The main lithologic units lie athwart the general direction of ice flow. The simplicity of the disposition of the lithologies on either side of the Moine Thrust belt (i.e. to the east of the Moine Thrust and to the west of the Sole Thrust) (Fig. 2.2) suggests the possibility of tracing long and well-defined erratics trains. The complicated repetition of lithologies in the Moine Thrust belt itself means that a study of erratics is only likely to show former ice-flow directions on a broad scale. However, these may be useful in isolating the postulated ice-shed zone.

Fortunately, a fairly large proportion of the study area is free of a thick blanket peat cover which would obscure erratics. Nevertheless, thick peat interrupted the mapping of erratics in several areas, notably in the Knockan basin and around Ledbeg.

Owing to the above advantages, it was hoped that such a study of erratics in the Assynt area would lead to a clearer understanding of the direction of former ice flow, the effect of an irregular topography on the dispersal patterns, and possibly the relative transportation distances of some of the lithologies.

4.2 RESEARCH METHODS

Prior to fieldwork the original six-inch Geological Survey map sheets of the Assynt area were studied and all comments pertaining to glacial erratics were noted on the author's field maps. Other indications of former ice-flow directions (*e.g.* striae and comments on ice-moulded bedrock forms) were also noted.

From the study of aerial photographs much of the field area appeared to be free of a drift cover. Therefore it was decided that the most appropriate erratic material to study would be that of boulder size (*i.e.* particles with diameters of greater than 200 mm). Blocks of such a size are much less likely to have been redistributed by fluvial processes since deposition, and are also readily located and identified in the field. The basic intention was to cover the whole of the area with repeated traverses across the direction of former ice flow (where topography would allow), to be carried out concurrently with the geomorphological mapping of the region. It was suspected that in the area of 'knock-and-lochan' topography west of the Sole Thrust a systematic survey of erratic boulders would reveal well-defined boulder trains from the Torridonian sandstone hills of Suilven, Canisp, Beinn Reidhe and Beinn Gharbh. Systematic sampling of the erratics was achieved by mapping those encountered along traverses which approximately followed the north-south 1 km grid-lines on the Ordnance Survey maps; after the boulder trains had been identified, this 1 km - spacing was reduced in certain instances to define more clearly the outer edges of each boulder train.

Before fieldwork commenced a preliminary study was made of hand specimens of some of the lithologies present in the Assynt area, held in the geological collections of the Grant Institute of Geology (University of Edinburgh) and in the Institute of Geological Sciences at Murchison House, Edinburgh. Various lithologies were also checked at their outcrops to ensure that erratic boulders could be correctly identified.

The following lithological types of erratic were used:

i) gneisses and other foliated rocks of the Lewisian complex,

ii) schistose rocks of the Moine group,

iii) grits and sandstones of the Torridonian group,

iv) Cambrian False-bedded quartzite and Pipe Rock (hereafter referred to collectively as 'quartzites'),

v) Canisp porphyry,

vi) Breabag porphyrite,

vii) various distinctive rock types from the Loch Borralan intrusive complex, including the pink quartz-syenites and the spotted borolanite.

These lithologies are easily identified, with the exception of Canisp porphyry and Breabag porphyrite which look very similar at first glance, although a closer examination can distinguish between them.

During fieldwork erratic boulders of the above lithologies were plotted on field maps at a scale of 1: 10,560. Later the data were transferred to maps of smaller scale so that the overall distribution could be more easily appreciated.

4.3 RESULTS

The results of the fieldwork are presented as a series of maps (Figs.4.1 to 4.6) showing the location of individual erratics, except where this is impractical owing to the abundance of mapped erratics. Before examining these spatial patterns it is necessary to isolate various factors that have influenced the mapped distribution pattern.

4.3.1 Factors influencing the plotted distribution pattern

A number of factors have probably influenced the distribution of glacial erratics in the Assynt area as presented in the figures. The distribution pattern has therefore to be interpreted with the following in mind, as well as those variables that affect the general dispersal of erratic material already listed in section 4.1.2.

i) The plotted distribution pattern reflects the amount of coverage of the area during fieldwork. This is partly controlled by the presence or absence of other geomorphic features that were being studied, and hence the amount of time spent in any one location, and also by the degree of accessibility of various parts of the field area.

ii) Certain areas are covered by deep peat which concealed any underlying erratics.

iii) The plotted distribution has also been affected by post-depositional processes of weathering, and redistribution by mass movement (especially by solifluction). Hence erratics of lithologies particularly susceptible to subaerial chemical and mechanical weathering may have been selectively destroyed, and erratics on certain slopes may have been moved to lower levels.

4.3.2 Glacial erratics and former ice flow patterns in the Assynt area

In order to interpret the distribution of glacial erratics shown in Figs. 4.1 to 4.6 in terms of the direction of former ice flow, certain general assumptions have to be made. Firstly it is assumed that the erratic blocks are largely representative of the last ice movements affecting that part of the field area in which they are found. Secondly, it has to be assumed that the outcrops of the various lithologies were the same in the past as those shown on the modern geological map. These assumptions are unlikely to be completely valid, as certain erratic boulders may have been transported to their present position by ice movements older than the last to have affected the area, and it is likely that glacial erosion has removed certain outcrops of a rock type, and reduced or enlarged others.

(a) Erratics from the Loch Borralan Complex (Fig. 4.1). These erratic boulders show a distribution predominantly westward and north-westward from the source area, although several boulders of distinctive borolanite were found on the eastern slopes of Cnoc na Glas Choille, south of the outcrop, along the line of a newly-cut Forestry Commission

Fig. 4.1 Distribution of erratics from the Loch Borralan complex and Breabag porphyrite outcrops.

Key: 1. Outcrops of the Loch Borralan complex.

2. Occurrences of Loch Borralan complex erratics.

3. Outcrops of Breabag porphyrite.

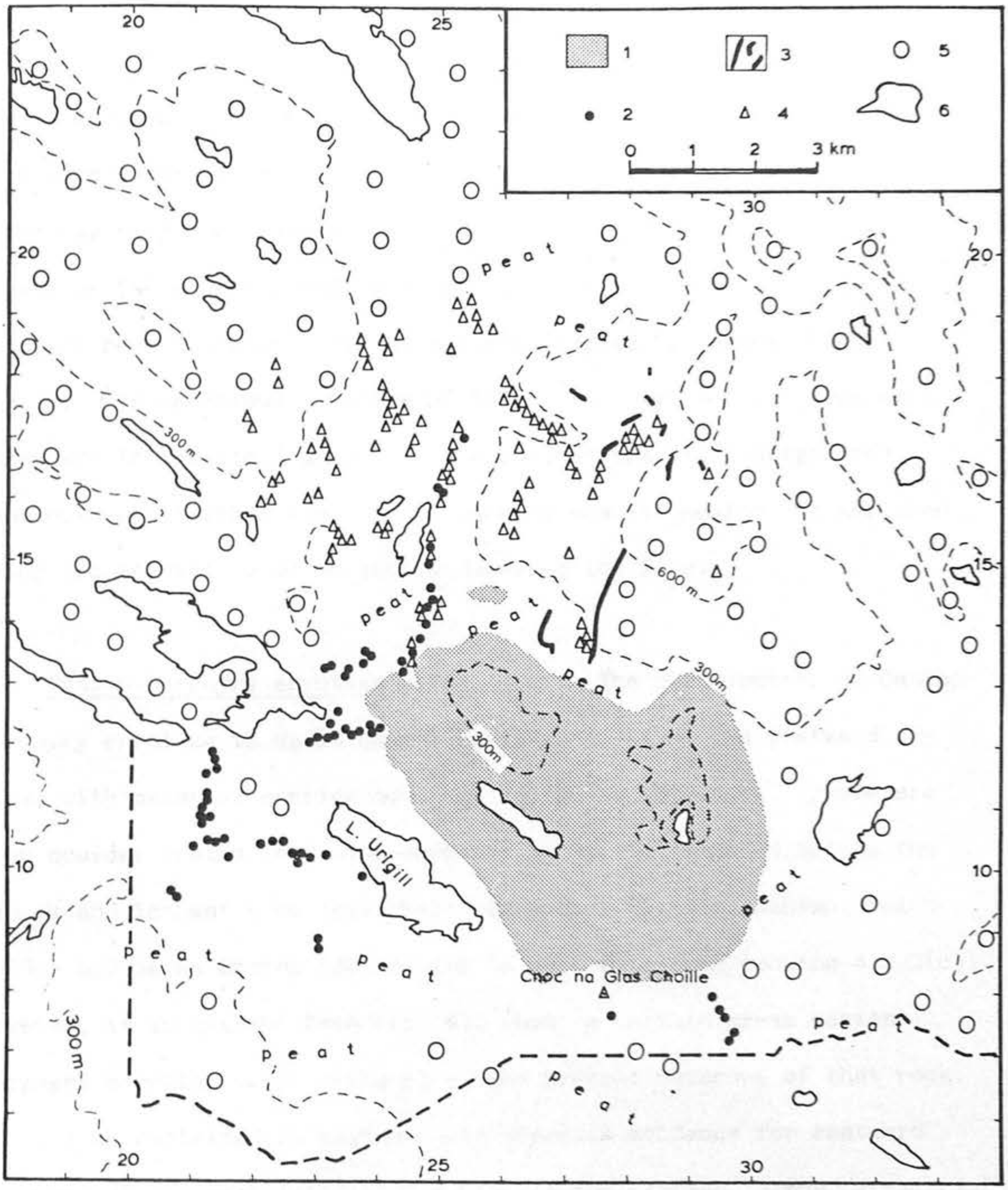
4. Occurrences of Breabag porphyrite erratics.

5. Areas examined where Loch Borralan complex erratics and Breabag porphyrite erratics were absent.

6. Main lakes.

Contours at 300 m intervals. (Heavy dashed line represents part of the boundary of the study area.)

(diagram overleaf)


1
2
3
4
5
6
0 1 2 3 km
20
25
30
15
10
peat
300 m
600 m
L. Urigill
Cnoc na Glas Choille

road. Unfortunately, the thick peat cover of the central part of the Knockan basin (south and south-east of Loch Urigill) has obscured the dispersal pattern in this area.

(b) Breabag porphyrite erratics (Fig. 4.1). Breabag porphyrite occurs only as sills in the Cambrian strata of the lowest dislocated rock mass in the Moine Thrust belt (Sabine 1953). Erratics of this lithology have been carried to between west and WNW : at only two locations (NC 29161627 and NC 29541471) have Breabag porphyrite erratics been found east of the outcrops mapped by Sabine (1953, Fig. 8). As anomalous erratics of this lithology have not been met elsewhere in the field area, it is suspected that they originated from outcrops farther east, now eroded by glacial action, rather than being representative of an eastward-moving ice flow.

(c) Canisp porphyry erratics (Fig. 4.2). The distribution of Canisp porphyry erratics is again on the whole indicative of a westward ice flow, with material carried at least as far as the coast. There are good boulder trains of Canisp porphyry erratics traceable across the 'knock-and-lochan' area from their sources on Suilven, Canisp, Beinn Reidhe and Beinn Gharbh (dealt with in more detail in section 4.3.3(c)). However, it is evident from Fig. 4.2 that in certain areas Canisp porphyry erratics occur eastward of the present outcrops of that rock. As will be explained in section 6.4, there is evidence for eastward movement of ice associated with a glacier that formed in the Lateglacial period in the area west and north-west of Stronchrubie. This can account for most of these anomalous erratics, but not all of them. Erratics near outcrop 'X' in Fig. 4.2 probably relate to old outcrops, farther east on the quartzite dipslope, which have been totally eroded

Fig 4.2 Distribution of Canisp porphyry erratics.

Key: 1. Outcrops of Canisp porphyry.

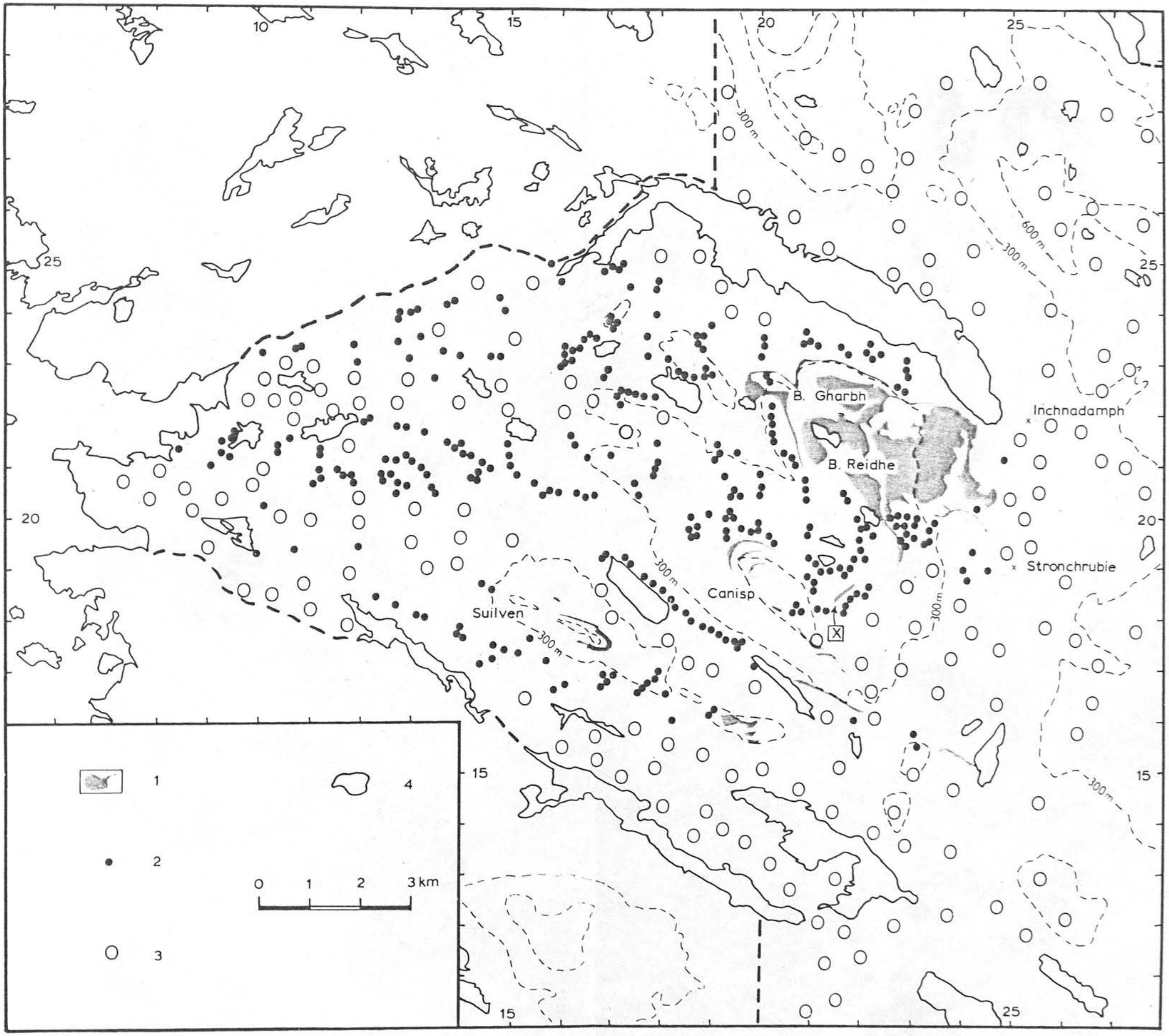
2. Occurrences of Canisp porphyry erratics.

3. Areas examined where Canisp porphyry erratics were absent.

4. Main lakes.

Contours at 300 m intervals.

(diagram overleaf)



10
15
20
25
25
20
15
15
15
25
300 m
600 m
B. Gharbh
B. Reidhe
Inchnadamph
Stronchrubie
Canisp
Suilven
X
1
2
3
4
0 1 2 3 km

by glacial action, but it is difficult to account for the Canisp porphyry erratic seen in the roadside cutting at NC 24852110 near Inchnadamph: this erratic lies outside the limits of the Lateglacial glacier mentioned above, and may have been brought eastward by melt-water, or by ice in an earlier period.

(d) Torridonian erratics (Fig. 4.3). Most of the Torridonian erratics have been found on the 'knock-and-lochan' area in the form of well-defined boulder trains from Suilven, Canisp and the Beinn Reidhe/ Beinn Gharbh area (see section 4.3.3(c)). Elsewhere, Torridonian erratics have a limited distribution. Blocks of sandstone on the eastern slopes of Canisp, in the Allt nan Uamh valley and on the plateau above Stronchrubie have apparently been carried westwards and north-westwards from the Torridonian masses of Beinn an Fhuarain and Beinn nan Cnaimhseag. Similar blocks east of Inchnadamph, including a boulder of basal conglomerate (at NC 26272176), appear to have been brought down the Traligill valley from a Torridonian outcrop on Conival. A solitary sandstone block at NC 21630913 may have been carried north-westwards from a Torridonian outcrop on one of the nappes near the Moine Thrust. A few sandstone blocks alongside the A437 road near Ledbeg might have been brought to the SSW from Beinn an Fhuarain, but a consideration of evidence from other erratics suggests that a more likely source is the Torridonian outcrop on the south-east slopes of Meall a' Bhraghaid. Torridonian erratics in Glen Oykel seem to have been transported down the valley by glacier ice. In view of the evidence from other erratics, it is likely that the solitary boulder of Torridonian conglomerate found well onto the Moine schist area between Meall a' Bhuirich and An Stuc (NC 34481052) came from the outcrop west of Dubh Loch Beag, or from the south-east

Fig. 4.3 Distribution of Torridonian erratics.

Key: 1. Outcrops of Torridonian lithologies.

2. Occurrences of Torridonian erratics.

3. Areas examined where Torridonian erratics were absent.

4. Main lakes.

Contours at 300 m intervals.

(diagram overleaf)

B. Gharbh
B. Reidhe
Canisp
Suilven
Inchnadamph
Stronchrubie
Allt nan Uamh
B. nan Cnaimhseag
B. an Fhuarain
Conival
Dubh Loch Beag
Meall a' Bhraghaid
Meall a' Bhuirich
An Stuc
Ledbeg
peat
300 m
600 m
1
2
3
4
0 1 2 3 km

slopes of Conival. This was the only Torridonian erratic found east of the Moine Thrust.

(e) Quartzite erratics (Fig. 4.4). Blocks of False-bedded quartzite and Pipe Rock occur over the whole of the area. As is apparent from Fig. 4.4, the outcrops of quartzite are repeated many times in the Moine Thrust belt, and occupy about half the total area of the 'Assynt window' (the embayment in the line of the Moine Thrust). For this reason quartzite erratics cannot supply information on detailed ice-flow patterns. Quartzite blocks have been carried westwards, supporting similar evidence from other, more localised, erratic lithologies. The more interesting fact from the plotted distribution, however, is that quartzite boulders have also been carried eastwards onto the area of Moine schist. Besides the quartzite erratics found on the Moine area during fieldwork, Fig. 4.4 also shows areas east of the study area where the Geological Survey reports their occurrence on the original six-inch geological map sheets, and in the local memoir (Read et al. 1926, p. 180). The author has seen quartzite erratics still farther east near Lairg railway station (NC 583039). The presence of quartzite boulders south and south-east of Loch Ailsh, on Cnoc Chaornaidh and An Stuc, suggests that they were transported here by ice issuing from Glen Oykel. The erratics on Cnoc na Sroine may also have been carried there by Oykel ice that turned and flowed westwards after rounding the eastern end of Meall a' Bhraghaid; alternatively, they may have originated from the quartzite outcrops of Creag Liath and Meall Diamhain, but such a south-western flow of ice is not supported by the distribution of other erratics.

(f) Lewisian erratics (Fig. 4.5). The distribution of boulders of

Fig. 4.4 Distribution of quartzite erratics.

Key: 1. Outcrops of Cambrian quartzite.

2. Occurrences of quartzite erratics.

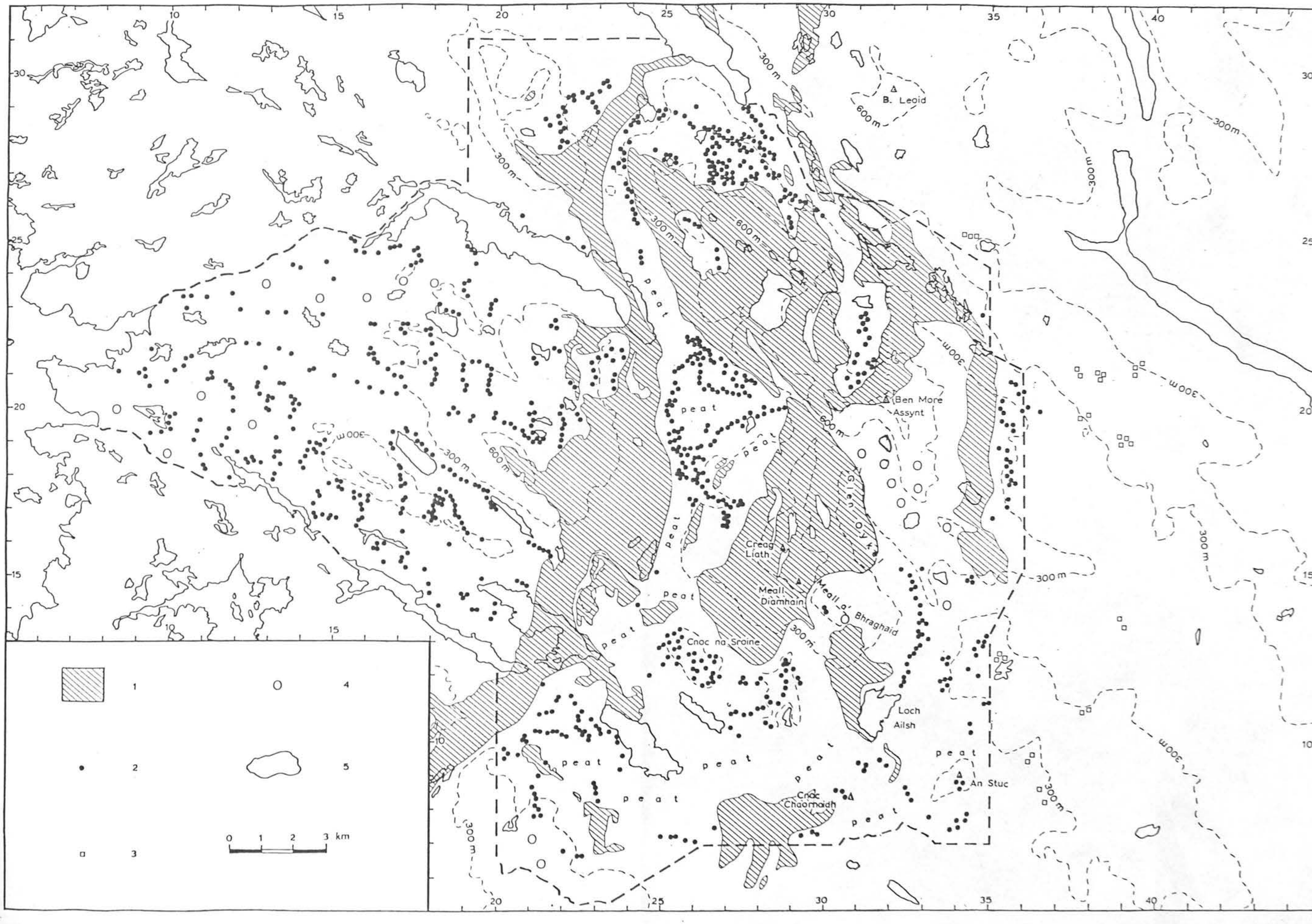
3. Occurrences of quartzite erratics noted by other workers.

4. Areas examined where quartzite erratics were absent.

5. Main lakes.

Contours at 300 m intervals.

(diagram overleaf)



B. Leoid
Ben More Assynt
Glen Oykel
Creag Liath
Meall Diamhain
Meall a' Bhraghaid
Cnoc na Sroine
Loch Ailsh
An Stuc
Cnoc Chaornaidh
peat
300 m
600 m
1
2
3
4
5
0 1 2 3 km

gneiss and other rocks from the Lewisian complex is one of the most important plotted. Fig. 4.5 shows that Lewisian rocks mainly occur in the eastern and north-eastern part of the Moine Thrust belt, and occupy all the ground west of the line of Torridonian sandstone hills from Suilven to Quinag. In the north of the area, Lewisian erratics have been transported north-westwards onto the quartzite dipslope of Spidean Coinich on Quinag, and over the col between Quinag and Glas Bheinn. Other Lewisian erratics emphasise the westward direction of former ice flow that has already become apparent from the distribution of previously described erratics of different lithologies. However, the most striking feature of the westward distribution of Lewisian erratics is that they have been carried onto the top of the main mountain ridge from Lewisian outcrops to the east. Hence Lewisian erratics occur on Beinn Uidhe, the northern part of Beinn an Fhurain, Breabag and Creag Liath, yet there are considerable areas on the ridge where Lewisian erratics have not been found : Glas Bheinn, the southern part of Bheinn an Fhurain, Conival and Ben More Assynt. Lewisian boulders have also been carried eastward onto the Moine area. Again, the distribution of erratics mapped in the field has been supplemented in Fig. 4.5 from various Geological Survey sources. Lewisian erratics have also been reported from the east side of Loch Shin, opposite the confluence with a tributary, the Allt Car (Read _et al_. 1926, p. 180); farther north-west, the author has noted Lewisian boulders near the Overscaig Hotel, again on the east side of Loch Shin. In many respects, therefore, the distribution of Lewisian erratics is similar to that of the boulders of quartzite. Hence the positioning of a former ice divide postulated to account for the dispersal of quartzite erratics both eastwards and westwards from the mountainous area must also account for the distribution of Lewisian

Fig. 4.5 Distribution of Lewisian erratics.

Key: 1. Outcrops of Lewisian rocks.

2. Occurrences of Lewisian erratics.

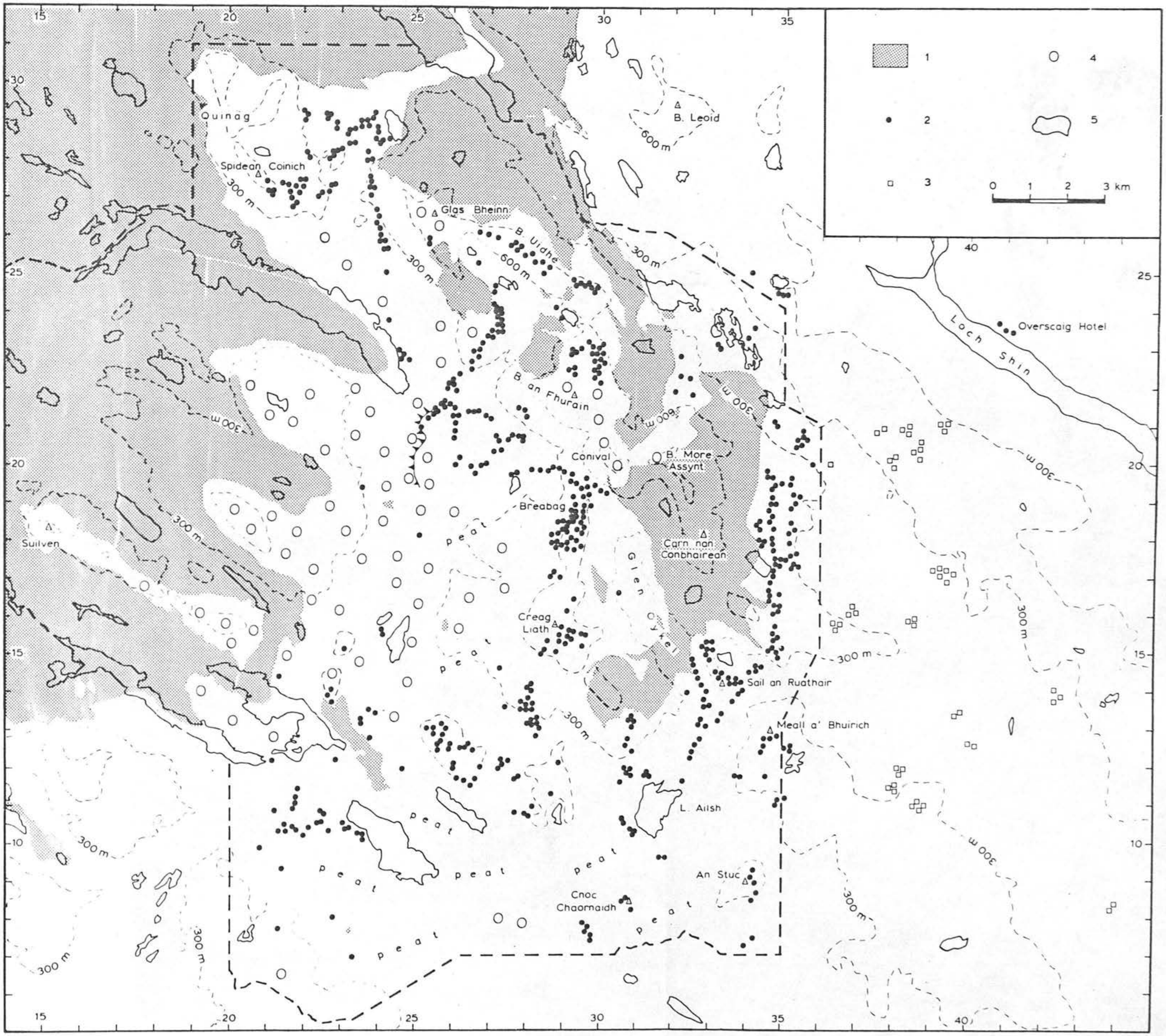
3. Occurrences of Lewisian erratics noted by other workers.

4. Areas examined where Lewisian erratics were absent.

5. Main lakes.

Contours at 300 m intervals.

(diagram overleaf)



Quinag
Spidean Coinich
Glas Bheinn
B. Uidhe
B. Leoid
B. an Fhurain
Conival
B. More Assynt
Breabag
Carn nan Conbhairean
Glen Oykel
Creag Liath
Sail an Ruathair
Meall a' Bhuirich
L. Ailsh
An Stuc
Cnoc Chaornaidh
Suilven
Loch Shin
Overscaig Hotel
peat
300 m
600 m
1
2
3
4
5
0 1 2 3 km

erratics.

Fig. 4.5 also shows that Lewisian erratics occur in the southern part of the 'Assynt window' where Lewisian outcrops are small, and on the hills of Cnoc Chaornaidh, An Stuc and Meall a' Bhuirich. It is suggested that these erratics were dispersed by ice flowing down Glen Oykel and spreading out south-westwards into the Knockan basin, and south-eastwards along the Oykel valley, as might have occurred at an early stage of the last glacial period. Subsequently, when the Assynt ice had coalesced with ice from the high ground to the south of the study area, the direction of ice flow in the southern part of the 'Assynt window' was mainly westward, redistributing the erratics originally brought southwards.

(g) Moine schist erratics (Fig. 4.6). There has been a limited dispersal of Moine erratics westward. Moine schist boulders are scattered all over the Knockan basin area (although the thick peat here may be obscuring a more continuous distribution pattern), and are abundant on Cnoc na Sroine, in the area west of Loch Ailsh, on the scattered quartzite slopes of Meall Diamhain and Creag Liath (Fig. 4.7a), and on the south-east slopes of Meall an Aonaich. North of this point Moine erratics have only been found in a narrow strip 1 km west of the Moine Thrust north of Loch Carn nan Conbhairean, and three isolated examples were found in the vicinity of Loch Bealach a' Mhadaidh. As the figure shows, no further instances of Moine erratics have been discovered despite a thorough search of the study area. The presence of Moine erratics west of the Moine Thrust must indicate that at some stage during the glaciation of the Assynt region the ice divide lay entirely over the Moine schist area.

Fig. 4.6 Distribution of Moine schist erratics.

Key: 1. Outcrop of Moine lithologies.

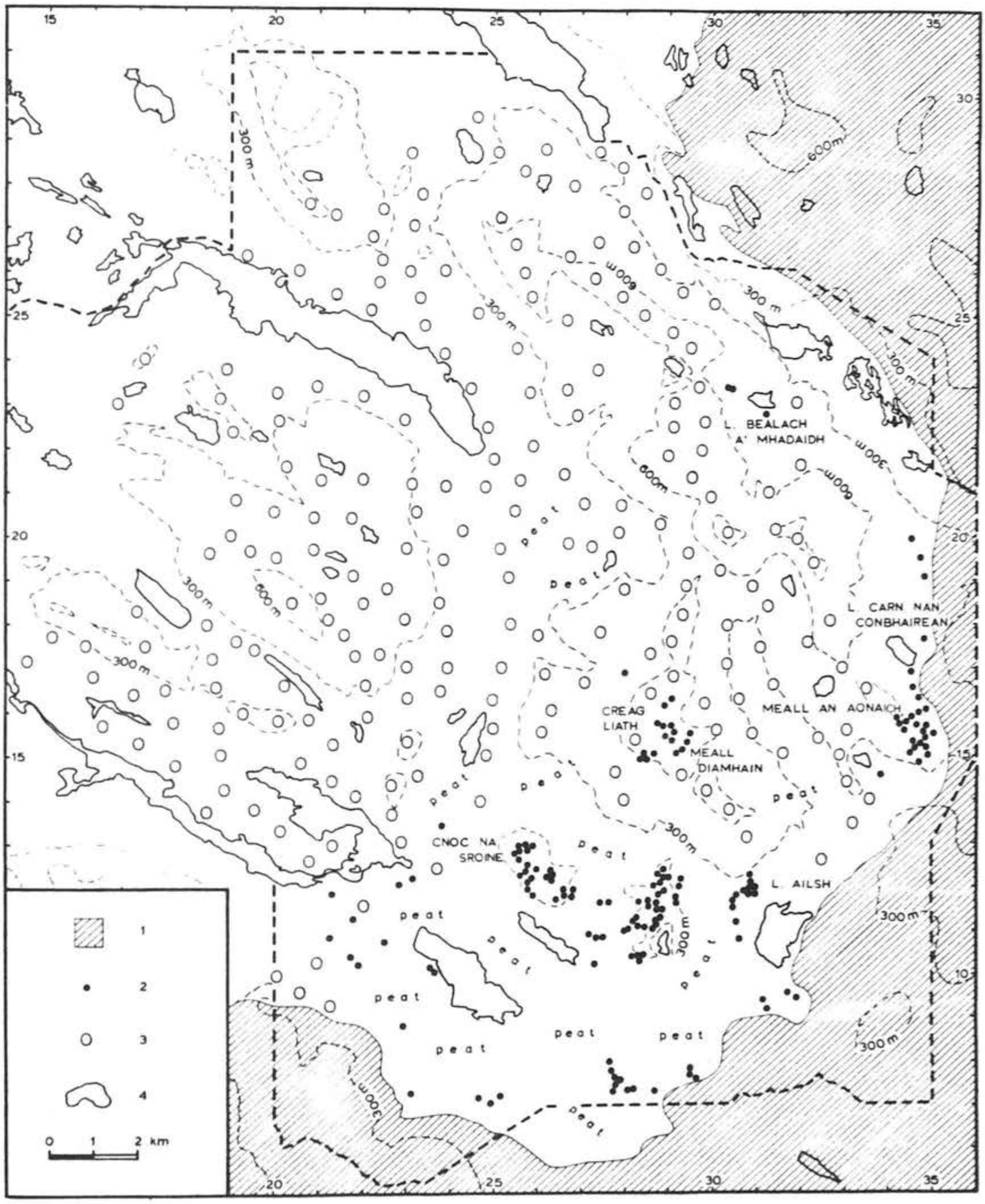
2. Occurrences of Moine schist erratics.

3. Areas examined where Moine schist erratics were absent.

4. Main lakes.

Contours at 300 m intervals.

(diagram overleaf)



L. BEALACH A' MHADAIDH
L. CARN NAN CONBHAIREAN
MEALL AN AONAICH
CREAG LIATH
MEALL DIAMHAIN
CNOC NA SROINE
L. AILSH
peat
300 m
600 m
1
2
3
4
0 1 2 km

The dispersal patterns of erratic boulders of various lithologies, described above, show that over most of the study area the last ice flow was generally towards the west or north-west. However, in the south-west of the area, the pattern is more complex : Lewisian erratics occur to the south and south-east of the main Lewisian outcrops, and Lewisian and quartzite erratics on the Moine schist area indicate an eastward ice flow whereas Moine erratics indicate one to the west. It is necessary to reconcile this apparently contradictory evidence.

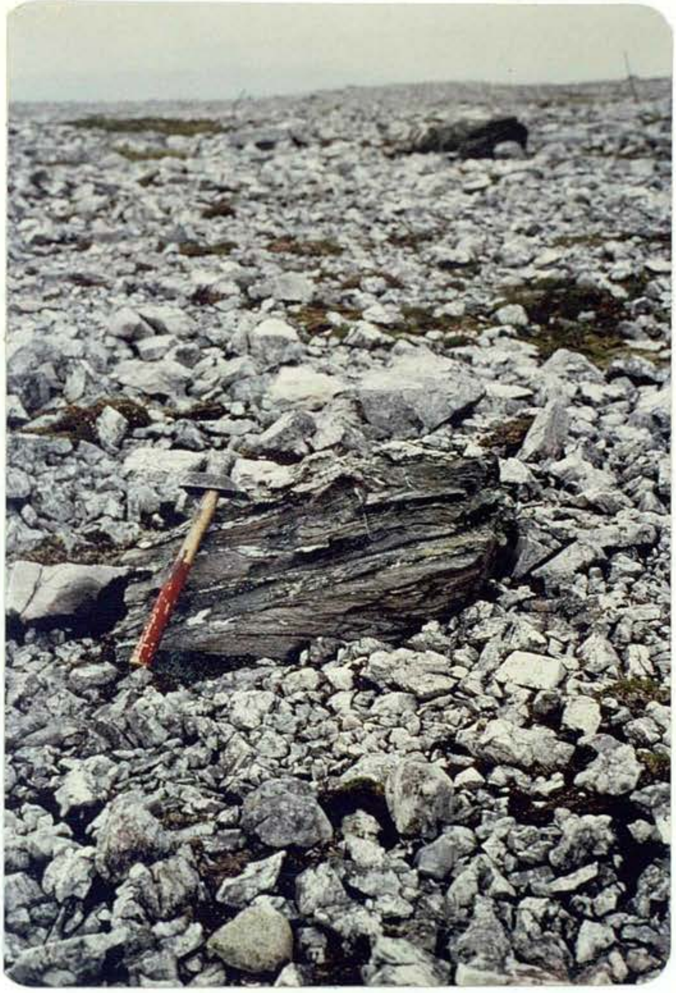
It is suggested, for reasons that will be discussed in detail in Chapter 6, that the onset of the last glacial period saw the development of glaciers in the mountain corries, notably in Glen Oykel and on the north and east side of the main mountainous ridge from Quinag to Meall an Aonaich. Such a development of glacier ice would explain the southward dispersal of Lewisian erratics, and the eastward dispersal of boulders of Lewisian gneiss and Cambrian quartzite onto the Moine schist outcrop. These glaciers continued to grow until the ice was thick enough to overtop the mountain ridge : at this stage an ice divide was established and ice flowed both eastwards and westwards.

This ice divide was always to the east of the Assynt mountains over the Moine schist area, though its exact position is a matter of some conjecture. However, the distributions of Moine and Lewisian erratics give useful clues. As has already been noted, Moine erratics occur solely in the southern part of the 'Assynt window', with only a few isolated examples found north of Ben More Assynt. On the mountain ridge farther north, only Lewisian erratics occur. Although the Lewisian erratics on Beinn Uidhe may have come from outcrops to the east and north-east, it is suggested that the

Fig. 4.7 Photographs: (a) Moine schist erratic on Creag Liath;

(b) Torridonian erratics from the Suilven boulder train (Suilven in the background).

(photographs overleaf)



a

b

ice flow over this part of the field area was to the WNW or NW, thereby sub-parallel to both the outcrop of the Moine Thrust plane in this vicinity and to the orientation of Beinn Uidhe. Such an ice-flow pattern can account for the mapped distribution of Lewisian erratics both on the mountain ridge and those carried north-westward onto the quartzite dipslope of Quinag. Moine boulders are restricted to those parts of the ridge in the south of the area where the more westward ice flow was not confronted by vertical rock faces. In the north of the area, the north-westward ice flow would have meant that few Moine boulders would have left the Moine outcrop (and Lateglacial glaciers (section 6.3) would have redistributed those that did).

4.3.3 Wider Inferences

Besides supplying information on the direction of former ice movement in the Assynt area, the distribution of erratics gives some important clues as to the characteristics of the last ice sheet, and the interaction of various ice masses.

(a) The distance of glacial transportation. It is interesting to note that of those that have been transported westwards, erratics from outcrops near to the postulated ice-divide zone (i.e. just east of the Moine Thrust in the eastern part of the field area) have been carried only short distances. For example, Lewisian erratics from the Ben More Assynt - Carn nan Conbhairean outcrop have only been carried westwards to the top of the Creag Sron Chrubaidh cliff near Inchnadamph, a total journey of about 6 km. Similar figures are obtained for the westward carry of erratics from the Loch Borralan complex and the Breabag porphyrite outcrops. Moine erratics have only been carried 9 km westwards. Such a restricted dispersal of erratics in the ice-shed zone is to be expected, as basal ice theoretically has a low velocity in such situations (e.g. Sugden & John 1976, Fig. 4.8a), increasing exponentially away from the ice divide until a maximum is reached beneath the equilibrium firn line, and thereafter decreasing towards the margin of the ice sheet. The eastward carry of erratics onto the Moine schist area does not match the restricted dispersal on the west side of the ice divide. Quartzite erratics have been transported at least 22 km to the vicinity of Lairg, and erratics of Lewisian gneiss have been carried eastwards at least 15 km. This can be explained by the ice having built up on the eastern side of the mountains, resulting in an eastward flow of ice (and hence eastward transportation of erratics) that lasted much longer than the westward flow.

Away from the ice-shed zone, quartzite, Torridonian sandstone and Canisp porphyry erratics have been carried from their respective source areas, on the hills of Suilven, Canisp, Beinn Reidhe and Beinn Gharbh, at least as far as the coast (some 13 km distant in the case of the quartzite erratics). This indicates that the ice sheet extended at least as far as the Minch, and probably much farther, as Torridonian sandstone erratics have been reported from Tolsta Head on north Lewis in the Outer Hebrides (von Weymarn 1979).

(b) The uplift of erratics by glacial ice in Assynt. The highest altitude attained by glacial erratics in the study area is over 750 m O.D., these being the Lewisian and Moine erratics found on Creag Liath. Lewisian erratics occur widely above 600 m on Breabag, Beinn an Fhurain, Beinn Uidhe, and Spidean Coinich. Although erratics have not been found on Conival or Ben More Assynt, blocks of ice-moulded quartzite showing striae and ice-moulding are present on the shattered quartzite ridge between them (at approximately 880 m O.D.), indicating that ice must have overtopped these mountains during one or more glaciations.

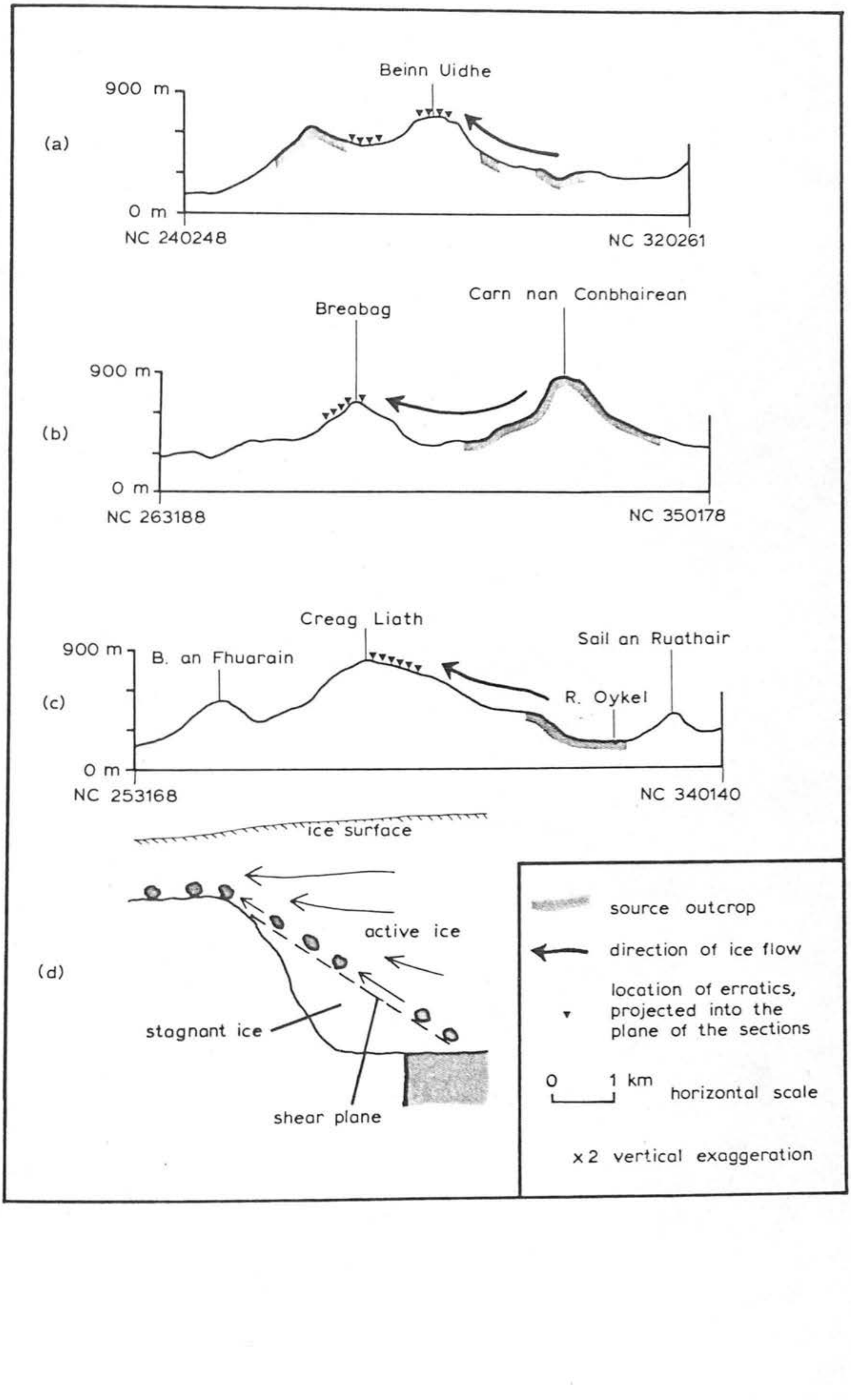
As has been mentioned previously in this chapter, it is possible to account for the uplift of glacial erratics to altitudes above their source outcrops as long as the ice surface is high above that of the highest relief features, and the slopes up which the ice flows are not very steep. Therefore, assuming a complete inundation of the topography by an ice sheet (i.e. its surface at greater than 1000 m O.D.), it does not appear difficult to explain the distribution of erratics on the eastern slopes of Canisp or Quinag, or on Cnoc na Sroine, all of which have relatively gentle proximal slopes. However, Lewisian erratics on Beinn Uidhe, Breabag and Creag Liath appear to have had to negotiate very steep east-facing slopes. Fig. 4.8 shows three profiles drawn across these mountains, sub-parallel to the direction of ice flow as shown by glacial striae, illustrating the outcrops of Lewisian gneiss and the places where Lewisian erratics have been found. The steepness of the proximal slopes is very apparent in the figure, even allowing for the x2 vertical exaggeration. Even under very thick ice, it is difficult to envisage the basal layers transporting large erratic blocks up free faces or other very steep slopes. In circumstances where flowing ice was confronted by such slopes, it is suggested that stagnation of a wedge of ice occurred, and high-angle shear planes developed between this stagnant ice and free-flowing ice above (Fig. 4.8(d)). Erratic blocks were carried over the stagnant ice wedge, to be deposited on the high ground on deglaciation. The gradient of such shear planes is likely to have been controlled by glaciological variables such as ice thickness, ice velocity and the temperature of the ice.

Fig. 4.8 The movement of glacial erratics up slopes.

(a), (b) and (c): selected cross-sections to illustrate the uplift of glacial erratics in the Assynt area from their source outcrops.

(c): suggested explanation of this uplift, involving movement along high-angled shear planes in the ice.

(diagram overleaf)


Beinn Uidhe
900 m
(a)
0 m
NC 240248
NC 320261
Carn nan Conbhairean
Breabag
900 m
(b)
0 m
NC 263188
NC 350178
Creag Liath
Sail an Ruathair
900 m
B. an Fhuarain
(c)
R. Oykel
0 m
NC 253168
NC 340140
ice surface
active ice
(d)
stagnant ice
shear plane
source outcrop
direction of ice flow
location of erratics, projected into the plane of the sections
0 1 km horizontal scale
x 2 vertical exaggeration

If one evokes such a process to account for the high-level dispersal of erratics shown in Fig. 4.8, it is possible to calculate the minimum and maximum straight-line gradients of the shear planes by relating the present altitude of the erratics to that of the nearest and the most distant part of the source outcrop. (In the case of the Lewisian erratics on Breabag (Fig. 4.8(b)), the minimum shear plane gradient was not calculated as the highest point of the Lewisian outcrop is higher than the altitude of the erratics.) The calculated shear plane gradients range from 1 in 2.0 to 1 in 6.2, equivalent to angles of 27° to 10° from the horizontal.

(c) Boulder trains. As has already been mentioned in section 4.3.2, well defined boulder trains of Torridonian grits and Canisp porphyry cross the 'knock-and-lochan' area (Figs. 4.2 and 4.3). The boulder trains are very important indicators of detailed ice-flow direction in this area, as glacial striae are few and the ice-moulded bedrock is largely structurally controlled. Definition of the margins of the boulder trains is relatively easy, as intervening areas are normally completely free of erratics, save for blocks of quartzite carried from the east which are present across the whole area. Fig. 4.9 outlines the limits of the boulder trains.

Two features that merit discussion are the shape of the boulder trains and their orientation. For ease of description the three main erratics trains have been named the Suilven, Canisp and Beinn Reidhe/Beinn Gharbh boulder trains. Other boulder trains run from Quinag in the north and Creagan Mor in the south, but both soon extend outside the study area and are therefore not considered here.

Each of the main boulder trains is long and thin rather

Fig. 4.9 The limits of the Assynt boulder trains.

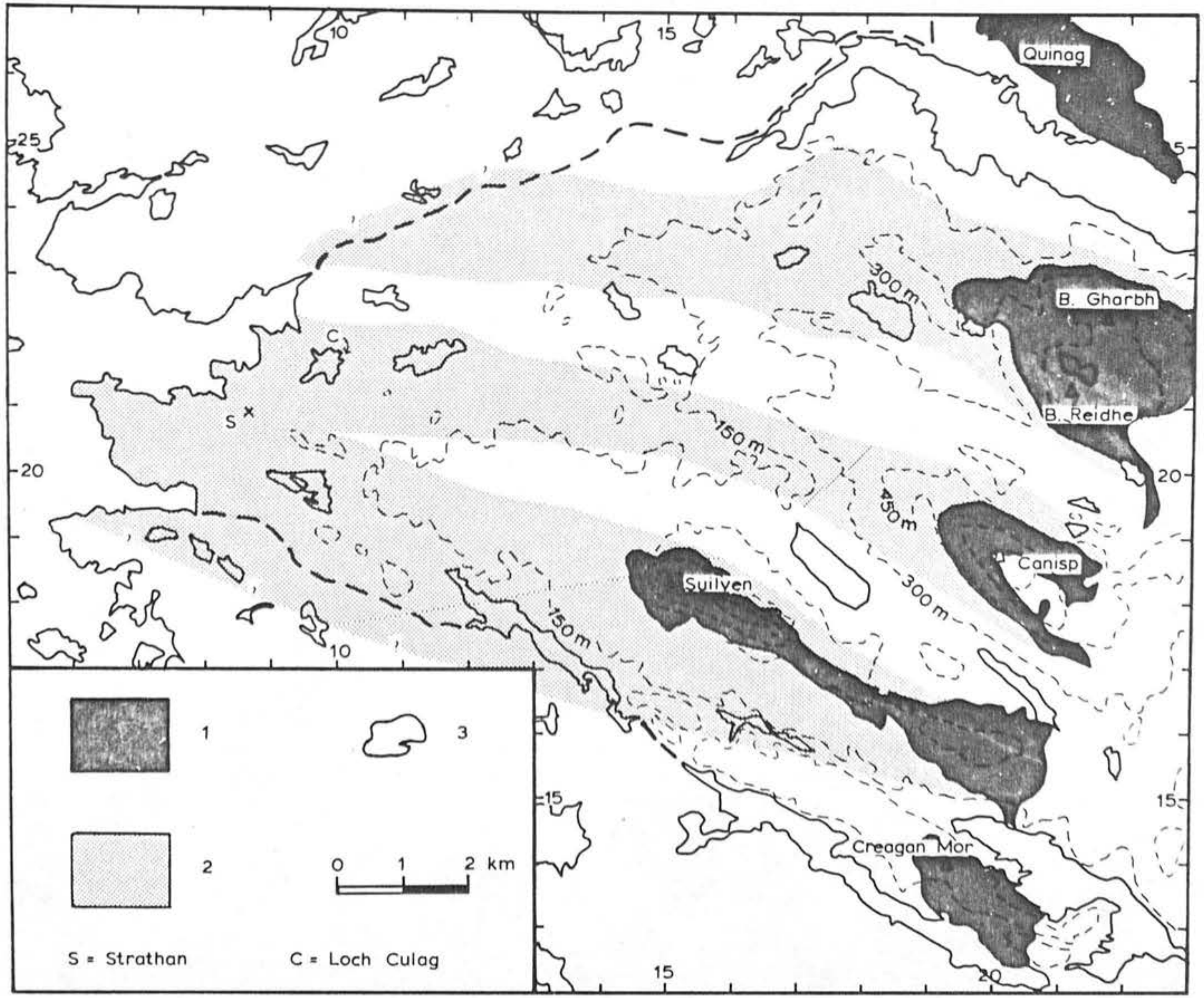
Key: 1. Outcrops of Torridonian rocks and Canisp porphyry.

2. Areas occupied by erratic blocks of the boulder trains.

3. Main lakes.

Contours at 150 m intervals.

(diagram overleaf)



Quinag
B. Gharbh
B. Reidhe
Canisp
Suilven
Creagan Mor
300 m
450 m
150 m
S
C
1
2
3
0 1 2 km
S = Strathan
C = Loch Culag

than noticeably fan-shaped. Both the Canisp and Beinn Reidhe/Beinn Gharbh boulder trains show a slight narrowing down-ice, whereas the Suilven boulder train widens slightly westwards. As divergence around topographic irregularities is often stated as being a major cause of widening erratics trains, this suggests that at least the smaller elements of the underlying topography were not markedly influencing the ice flow. This in turn implies that the distribution of these boulders relates to ice movements under a thick ice sheet. Further evidence to reinforce this point is given by the orientation of the Suilven boulder train, which follows a more westward azimuth than the orientation of Suilven itself, despite the apparently strong streamlining of the latter. It is therefore probable that Suilven was covered by a considerable thickness of ice at the time the present boulder train was formed; other boulder trains from this mountain relating to earlier phases of the last ice sheet may well have had quite different orientations, but such evidence, if it existed, has been removed by subsequent ice flow.

Another interesting feature of the boulder trains is that, while the Suilven boulder train appears to maintain the same orientation throughout its length (although its southern boundary has not been mapped as it lies outside the study area), the two northern boulder trains show a gradual change of direction between eastings 14 and 16, to the west of which they follow a more westward course. The three boulder trains therefore appear to converge on one another : indeed, the Canisp and Suilven boulder trains seem to merge in the vicinity of Strathan and Loch Culag. This curve in the line of the boulder trains to some extent mirrors a similar flexure in the grain of the topography; the marked south-westward bend at the western end of Loch Assynt also occurs at about the same longitude. Unfortunately it is impossible to say whether the topography is the cause or effect of this ice-flow pattern. However, the apparent convergence of ice in this region may be due to the deflection of ice around the southern end of the Quinag ridge, as even under a thick ice sheet the presence of large mountain masses is likely to have influenced the pattern of ice flow.

4.4 CONCLUSIONS

A study of the distribution of erratic boulders of seven different lithologies has given information about the glaciation of the Assynt area. The erratics show that the area has been completely glaciated by an ice sheet extending at least as far as the coast, and probably much farther at the glacial maximum. At the height of the last glaciation, the ice sheet was very thick, and probably covered the highest ground to a considerable depth.

An explanation of an apparently anomalous distribution pattern, whereby Lewisian and quartzite erratics have been carried eastwards and Moine erratics westwards in the same area, involves the initial development of glaciers in the corries on the eastern side of the main mountain ridge; thickening of these glaciers and further eastward development continued until such a time as the ice overtopped the mountain ridge and flowed both eastwards and westwards as an ice divide was formed. This ice divide always lay to the east of the Assynt mountains, over the Moine schist outcrop. The development of a southward-flowing valley glacier in Glen Oykel at an early stage of the glaciation can account for the presence of Lewisian erratics in areas to the south of the Lewisian outcrops.

The distance of transportation of erratics from various sources seems to support current views on basal velocities beneath ice sheets, in that the erratics have been carried a limited distance in the ice-shed zone. An explanation of another phenomenon in areas glaciated by former ice sheets, namely the high-level distribution of erratics that seem to have been transported up very steep rock surfaces, involves the development of shearing between lower static ice and upper moving ice. Such shear planes are shown to have been relatively high-angled features.

CHAPTER 5: THE DIRECTION OF FORMER ICE MOVEMENT: (b) GLACIAL STRIATIONS

5.1 INTRODUCTION

5.1.1 Glacial striations

Glacial striations, or striae, are linear or curvilinear scratches effected by abrasion at the base of a glacier. They are found on bedrock crossed by the ice and on transported stones. Agassiz (1838) was the first to assign a glacial origin to these features. Subsequent work by others showed that their distribution was widespread in areas previously covered by ice sheets; of these early reports, the monograph by T.C. Chamberlain (1888) gave by far the most detailed description of the various forms of glacial striae and their distribution on glaciated surfaces.

Typically, glacial striae are only a few millimetres in depth and several centimetres to tens of centimetres in length. However, there is a size continuum from micro-striae, invisible to the naked eye, up to giant grooves such as those described by Smith (1948). Laverdière & Guimont (1975) discussed the terminology of the various striation features at different scales.

Chamberlain (1888) described instances of striations occurring on all forms of rock surfaces : glacial striae were recorded on level surfaces, on surfaces both sloping with and sloping against the direction of ice flow, and oblique to such surfaces. Glacial striations were also noted on sheer, upright rock faces, incised both horizontally and vertically, and on overhanging surfaces in certain situations. Similar observations were made by Dort (1937) and Demorest (1938). Glacial striations are usually considered to be most prolific on gently-inclined surfaces that the ice was forced to ascend. However, the observed distribution of striae in an area is likely to be affected by several factors. Chamberlain (1888, pp. 158-160) discussed the following variables:

i) the relative distributions of glaciated bare-rock surfaces and those covered with surficial deposits,

ii) the degree of post-glacial preservation,

iii) bias introduced by the varying intensities of search for such features by the field worker,

iv) the interaction of topographic controls and glaciological variables,

v) spatial variations in the bedrock, especially variations in hardness.

Charlesworth (1957, p. 246) noted that rocks of a coarse texture, such as granites or grits, only took the bigger and deeper striations and grooves, whereas the tougher fine-grained rocks possessed delicate striations. Mineralogically soft, fine-grained rocks such as limestone are finely striated immediately after glaciation, but weathering quickly removes such features.

Striated bedrock can be produced by other agents of erosion than glacier ice. Dyson (1937) described striations caused by a snowslide, and Dionne (1973a) discussed the differences between glacial striae ('stries glaciaires') and those produced by floating ice such as icebergs ('stries glacielles'). Charlesworth (1957, p. 246) listed other agents capable of producing striations as "buffaloes or other animals, boulders trailed by seaweed, mountain torrents, winds and landslips, partially consolidated lavas gliding over one another, creep, mudflows or solifluxion, ... (and) tectonic movements", to which can be added volcanic 'nuées ardentes' (Flint 1957, p. 58) and scratches produced by heavy machinery, etc. (Thwaites 1956, p. 23). These can generally be isolated from glacial striations by a study of their distribution and orientation with respect to local slopes and other features.

5.1.2 Glacial abrasion

Glacial abrasion is due to rock fragments being forced against, and dragged along, the glacier bed by sliding ice. Striations are formed as part of this abrasion process "by the motion of rock asperities that locally deform and fracture the rock of the glacier bed and remove the debris" (Hallet 1979, p. 40). If the significance of the presence of striated bedrock in an area is to be properly discussed, it is necessary to understand subglacial abrasion processes.

Zumberge (1955) saw the factors controlling subglacial erosion (both abrasion and 'plucking') as falling into two categories, one pertaining to the applied force and the other to the resisting force. He saw ice thickness, the velocity and direction of ice flow, the amount and kind of basal debris, and the temperature regime of the basal ice as factors determining the applied force; all are inherent in the abrasion process itself. Factors determining the resisting force, and hence not directly responsible for the manner in which the erosion process is applied, but rather accounting for the manner in which the process manifests itself, were seen as variations in the topography of the ground surface, variations in rock structure and variations in lithology.

Recent textbooks that discuss subglacial processes (e.g. Sugden & John 1976; Embleton & Thornes 1979; Derbyshire et al. 1980) concentrate on the quantitative and theoretical work of G.S. Boulton and co-workers (especially Boulton 1974, 1979; Boulton and Vivian 1973). In cold glaciers (i.e. those whose basal layers are frozen to the bedrock) basal sliding is absent, and therefore abrasion is limited to those boulders and rock humps that protude into the zone in which shearing is concentrated. Abrasion may be totally absent where cold glaciers possess no basal debris. However, under temperate ice the presence of basal sliding means that abrasion is much more widespread.

Potential abrasional tools, in the form of rock particles of varying sizes, can have one of two sources : debris falling onto the surface of glaciers in their accumulation areas reaches the sole of the glacier along basal flow lines (Boulton 1978), or rock particles can be plucked from the bedrock that the ice overrides. Effective normal pressure to keep the rock particles in contact with the bed is provided by the interplay between the weight of ice above the particle (determined by the thickness of ice) and the degree of buoyancy provided by basal meltwaters under pressure. Basal meltwaters also lubricate the sole of the glacier, thus facilitating basal sliding, and to a certain extent remove abrasion products and hence prevent the build-up of a layer of rock flour that might inhibit further abrasion. The frictional drag between a basal rock particle and the bedrock is the product of the effective normal pressure, the area of contact between the bed and the particle, and a coefficient of dynamic friction (Boulton 1974). As the larger bedrock obstacles and coarser rock particles bear the higher shear stresses, very powerful abrading tools are provided by large blocks with small areas of contact with the bed, or by small particles trapped beneath large ones. Basal debris particles may be dragged or rolled across the glacier bed if the adhesion between the particles and the ice exceeds the force of internal friction in the subjacent rock fragments.

The rate of abrasion for a given basal sliding velocity theoretically increases with increasing effective normal pressure until a critical value of the latter is reached, beyond which plastic deformation of the ice around the particle will occur, as friction between the particle and the bed increases. An increase in effective normal pressure beyond this limiting value corresponds with a decrease in abrasion rate, until the frictional force exceeds the tractive force and abrasion ceases as the particle stops moving.

Besides these purely glaciological controls, the characteristics of the bedrock and the basal rock debris must be taken into account. The importance of various shapes and sizes of rock particles has already been alluded to, but of equal importance is the permeability of the bedrock (and its subsequent effects on effective normal pressures due to the presence or absence of basal meltwaters), its liability to fracturing, and the relative hardness of the bedrock and the abrading particles.

In the Boulton model of glacial abrasion, normal pressures at the base of the glacier, determined by ice thickness, are considered to be an important factor in the abrasion process. Recently, Hallet (1979) has argued that basal pressure (i.e. normal pressure) is not an important factor : subglacial rock particles will probably be completely surrounded by pressurised water or ice except where they touch the bedrock, and therefore basal pressures exert equal stresses to contact areas and to adjacent zones. "The effective force of contact is then independent of basal pressure and, hence, of the glacier thickness; it is only dependent on the buoyant weight of rock particles and on the viscous drag induced by ice flow towards the glacier bed due to basal melting and longitudinal straining of basal ice" (Hallet 1979, p. 40). Hallet admitted that his model simulates only those temperate glaciers whose basal layers contain merely a few rock fragments. This model cannot simulate abrasion beneath temperate glaciers with a large basal load, such as those described by Boulton (1974), as it does not take into account particle interactions and the effect of debris on basal sliding. "For particular rock types and concentrations of rock particles in the ice, calculations of the abrasion-rates reduce to evaluations of the effective force pressing particles against the bed and of the particle velocity" (Hallet 1979, p. 40). Hallet concluded that if all other parameters are equal, abrasion will tend to be fastest where basal melting is most rapid. As basal melting is related in part to ice thickness, Hallet's conclusion that abrasion is not greatly affected by glacier thickness must be viewed somewhat sceptically.

In summary, it is apparent that certain factors influence the efficacy of glacial abrasion processes and therefore control to a large extent the distribution of abrasion features such as glacial striae. Sugden & John (1976, pp. 153-156) have conveniently listed these factors. The presence of basal debris, sliding basal ice and processes transporting the debris down towards the bedrock, renewing the abrasives, are all seen as fundamental requirements for glacial abrasion. Several other factors affect the rate and type of abrasion : the thickness of glacial ice, the basal water pressure, the size and shape of entrained rock particles, the relative hardness of the rock particles and the glacier bed, and the degree of removal of rock flour and other abrasion products by meltwater. It will be necessary to discuss the significance of the striae mapped in the Assynt area in terms of the above.

5.1.3 Striae and the direction of ice flow

Any glacial striation can have resulted from two diametrically opposed directions of ice flow (Flint 1957, p. 58; Embleton & King 1975, p. 183). Certain earlier workers attempted to determine the direction of ice flow from the form of the striae (e.g. Charlesworth 1957, p. 247). It was suggested that basal rock particles tended to abrade bedrock surfaces gradually at first, the striation steadily deepening and broadening as the abrading point was ground away, and often terminating abruptly; striations of such a form were referred to as 'nailhead' striae, with their wider ends pointing down-ice. However, forms which are exactly the reverse could be produced by a rock particle brought abruptly and forcibly into contact with the bed, and gradually being withdrawn into the ice. In fact striae exhibit a multitude of different forms when closely inspected, as illustrated by Chamberlain (1888, pp. 229-230), and it would be very unwise to attempt to determine the direction of ice flow purely from the physical attributes of certain of the striae.

For this reason, most workers in this field stress the need to consider all the available alternative lines of evidence, in conjunction with the striation trends, to ascertain the direction of ice movement across a region. Hence certain Scandinavian workers have looked at the distribution of erratics in their study areas, analysed till fabrics, and looked closely at the ice-moulding of bedrock to supplement information gathered from an analysis of striation orientations (e.g. Edelman 1951; Virkkala 1951, 1960; Johansson 1968; Sindre 1974).

In many instances more than one set of striation trends is discovered. Sometimes bedrock exposures possess two or several striae trends, some actually crossing. In these cases, it is obviously as important to determine the relative ages of the trends as it is to ascertain from what directions the ice that formed them was flowing. Edelman (1951) noticed that certain roche moutonnée forms were striated obliquely to the orientation of the moulded rock. Hence he concluded that the obliquely-trending striae represented a younger abrasion period than that which formed the roche moutonnée, which may be represented elsewhere by a set of striae. In addition, he noticed that in certain cases facets cut into the roche moutonnée possessed older striae of yet another trend. Such facets are most likely to be preserved in lee-side positions (Fig. 5.1). Virkkala (1951, 1960) noted similar relationships between striations and their positions on ice-moulded rock surfaces, but in situations where no such evidence was available he argued that when sharper, finer striae crossed larger, coarser ones, the latter were older, having survived from an earlier phase of abrasion. As a general rule, this relationship seems to be a fairly good guideline, but it cannot be accepted with complete confidence unless the finer striae are actually seen to cut into the bottom of the coarser ones (Fig. 5.2). Otherwise it is possible to take an extreme view, and argue that the more pronounced, coarser striae are the most recent, abrasion having removed the upper portions of those now represented by finer striations. It is often impossible in the field to ascertain which set of striae cuts another : in such cases it is often useful to take a cast of the crossing striae and examine them microscopically under laboratory conditions (Svensson 1957; Markgren & Frisen 1963). Where two distinctly different trends of striae are found in an area, it may be inferred that the set that appears to be more widespread and better represented is likely to be the more recent. However, an older set of striae might be protected by a cover of thick drift during a later glacial phase and therefore may be better

Fig. 5.1 The development of striation sets of different ages on certain roches moutonnées (after Edelman (1951)).

1. Original roche moutonnée, with striated surface, formed by ice moving from the north.

2. Subsequent ice flow from east of north reshapes roche moutonnée, except for one facet of the old landform containing old set of striae.

3. Final ice flow from the north-east replaces the striae of stage 2, but abrasion is not sufficient to reshape roche moutonnée nor to totally remove striae of stage 1 protected on leeward side.

(diagram overleaf)

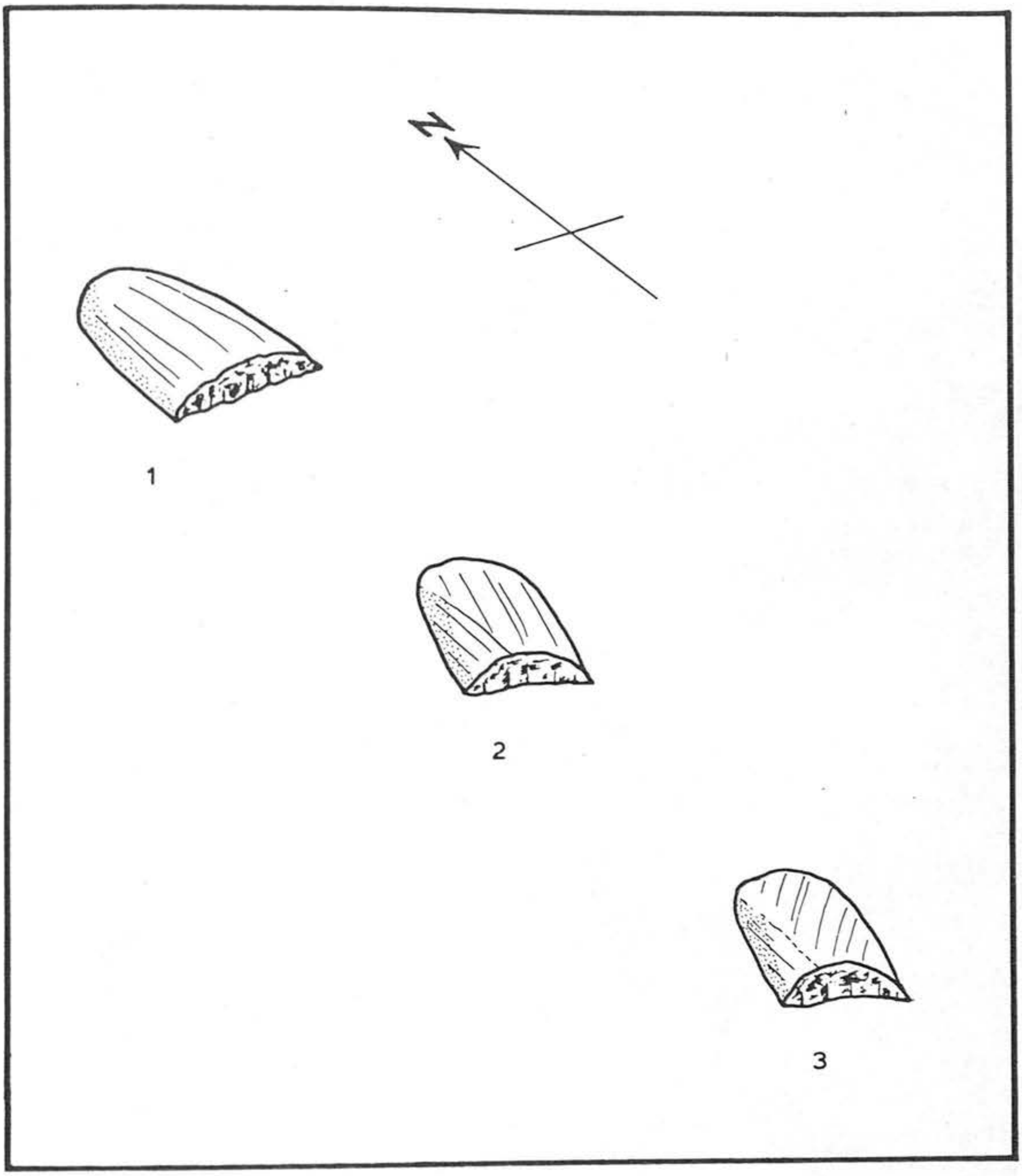
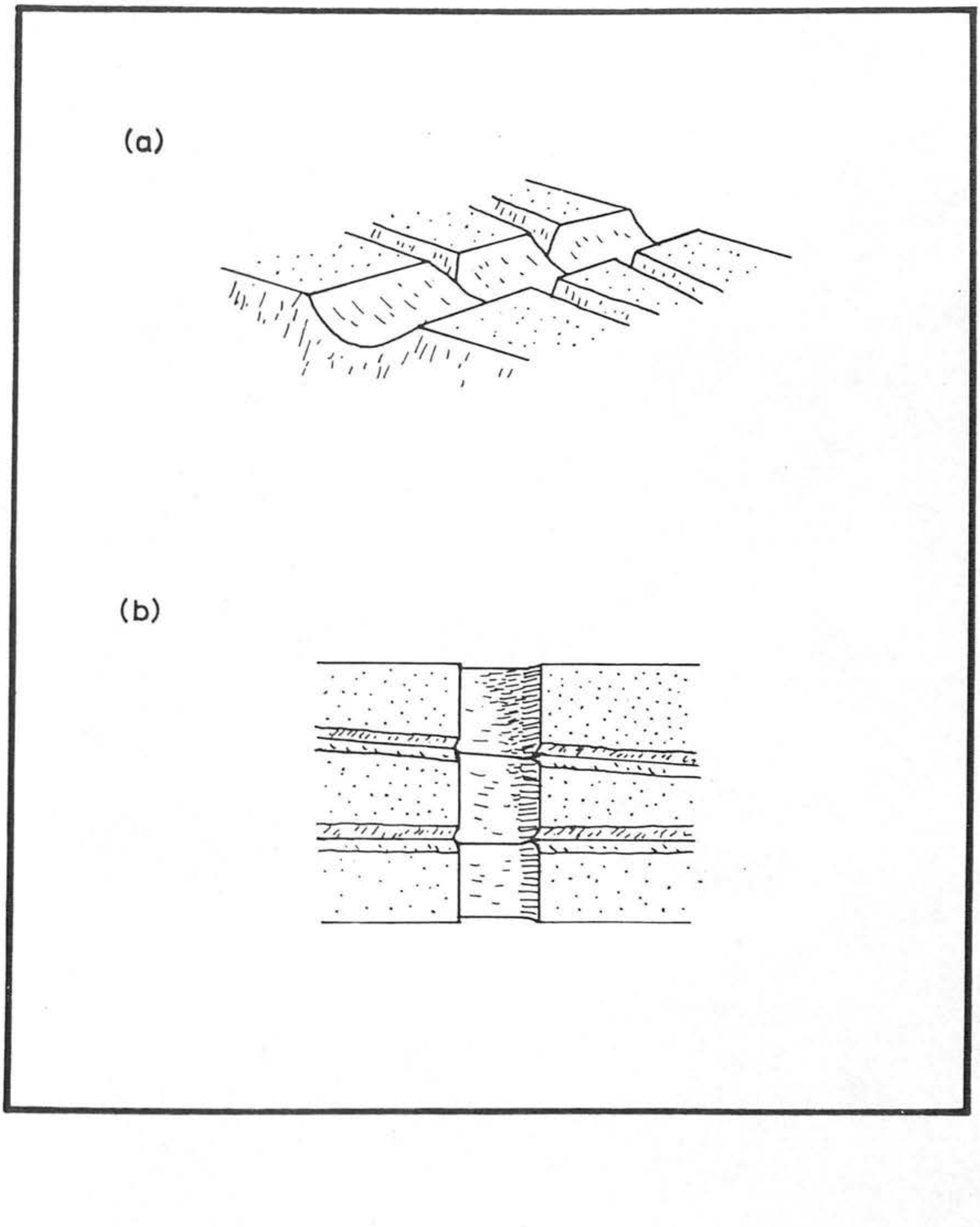

N
1
2
3

Fig. 5.2 Distinguishing the relative ages of two superimposed sets of glacial striae.

(a) oblique view and (b) plan view. The thinner striae cut across and into the larger striation, and are therefore the younger of the two sets.

(diagram overleaf)


(a)
(b)

represented in an area when postglacial erosion or mass movement reveals the old surface; in such an instance the old set of striae is only likely to be better represented locally.

Interglacial and interstadial weathering is likely to remove much evidence of previous glacial phases, and a subsequent ice advance will re-abrade rock surfaces, hence reorientating striation patterns; it is therefore unlikely that different sets of striations represent successive glaciations. They are more likely to represent different phases of the same glaciation, due to a shift in the position of ice dispersal centres, variations in the pattern of ice flow at glacier margins, or the varying influence of topographic controls exerted on a thinning glacier or ice sheet during deglaciation (Flint 1957, pp. 61-62).

Any single striation on a rock surface represents only the trend of the former local ice flow at that point. The ice flow can be deflected by topographic irregularities both at the large and small scales. Hence the distribution and orientation of hills and valleys has a profound effect on ice-flow directions, especially when the ice is thin. Even microtopographic irregularities can cause local divergence and deflection in the pattern of flow : Holmes (1937) reported a 40° deflection of ice moving into a small channel formed along a joint in the bedrock, and Edelman (1951) and Dreimanis (1956) described divergences of 50° and 72° respectively around ice-moulded bedrock humps. Andersen & Sollid (1971, Fig. 13) illustrated the typical striation pattern around the stoss end of a roche moutonnée. Therefore if regional trends of former ice movement are to be detected from observations of glacial striae, the orientation of a large number of them must be measured at various sites and over a large area, care being taken to measure striation trends where they have been least affected by local topographic irregularities (e.g. on the crests of roches moutonnées).

5.2 THE MAPPING OF STRIATIONS IN THE ASSYNT AREA

The only other attempt at mapping glacial striations in the Assynt area was undertaken by the Geological Survey at the turn of the century. Therefore, all the striations shown on the original 1 : 10,560 geological map sheets were plotted onto field maps at a similar scale. However, when checked in the field it was decided that the Geological Survey striae were not sufficiently accurate. As striated bedrock was widespread, the Geological Survey striae were subsequently ignored.

When a glaciated surface was encountered during fieldwork, that part of the surface where ice flow would have been least affected by the local topography (usually at the crest of ice-moulded rock protuberances) was examined for striations. Usually striae varying in direction by 5-15° were found on any one surface. An average orientation value of 10-20 striae was measured at each site to the nearest 1° from magnetic north, using an oil-filled type 3 Silva compass. Fairly strong orientations usually prevailed at each site, especially on quartzite bedrock. The mapped striae were eventually transferred to 1 : 25,000 map sheets, the measured directions being adjusted to grid north.

5.3 RESULTS

The orientations of striae were measured at over 600 locations during the course of fieldwork. The results are presented as Fig. 5.3. Due to the small scale of the figure, certain of the sites have had to be omitted, these being only sites whose omission does not detract from the striation pattern. Greater detail of striations in certain areas is given in Figs. 5.4, 5.5 amd 5.6. Arrowheads denote the probable direction of ice flow, as indicated by ice-moulded bedrock and the distribution of glacial erratics. In areas where doubt remains only the trend is shown.

It can be seen from the figure that the striae make up a logical pattern across the study area, paying scant regard to local slope directions. For this reason it can be reasonably inferred that they were formed by subglacial abrasion processes rather than by non-glacial agents such as those listed in section 5.1.1.

5.3.1 Factors influencing the distribution pattern

The distribution pattern that emerges in Fig. 5.3 has to be considered in relation to various factors. The most important factors in this case are geological. The petrological differences between the various rocks of the Assynt area, and the way in which they tend to crop out perpendicular to the main trend of former ice movement (largely due to the geological structure of the region), has meant that the ice flowed across rocks of varying relative hardness. The harder lithologies provided very effective tools for subglacial abrasion, and the striated rock surfaces are best preserved on those less prone to chemical and mechanical attack. There is a fairly close correlation between the distribution of striae (Fig. 5.3) and the distribution of Cambrian quartzite (Fig. 4.4). The quartzite is very hard due to its lithological composition and homogeneity, and surface features are well preserved as it is chemically inert. However, severe frost action can

Fig. 5.3 The distribution of glacial striations in the Assynt area.

Key:
1. Glacial striae, showing direction of former ice flow (determined from other evidence).
2. Glacial striae trend where directional indicators were absent.
3. Free faces.
4. Main lakes.
5. Inset areas shown in greater detail in other figures.
6. Contours at 300 m intervals.

(diagram overleaf)

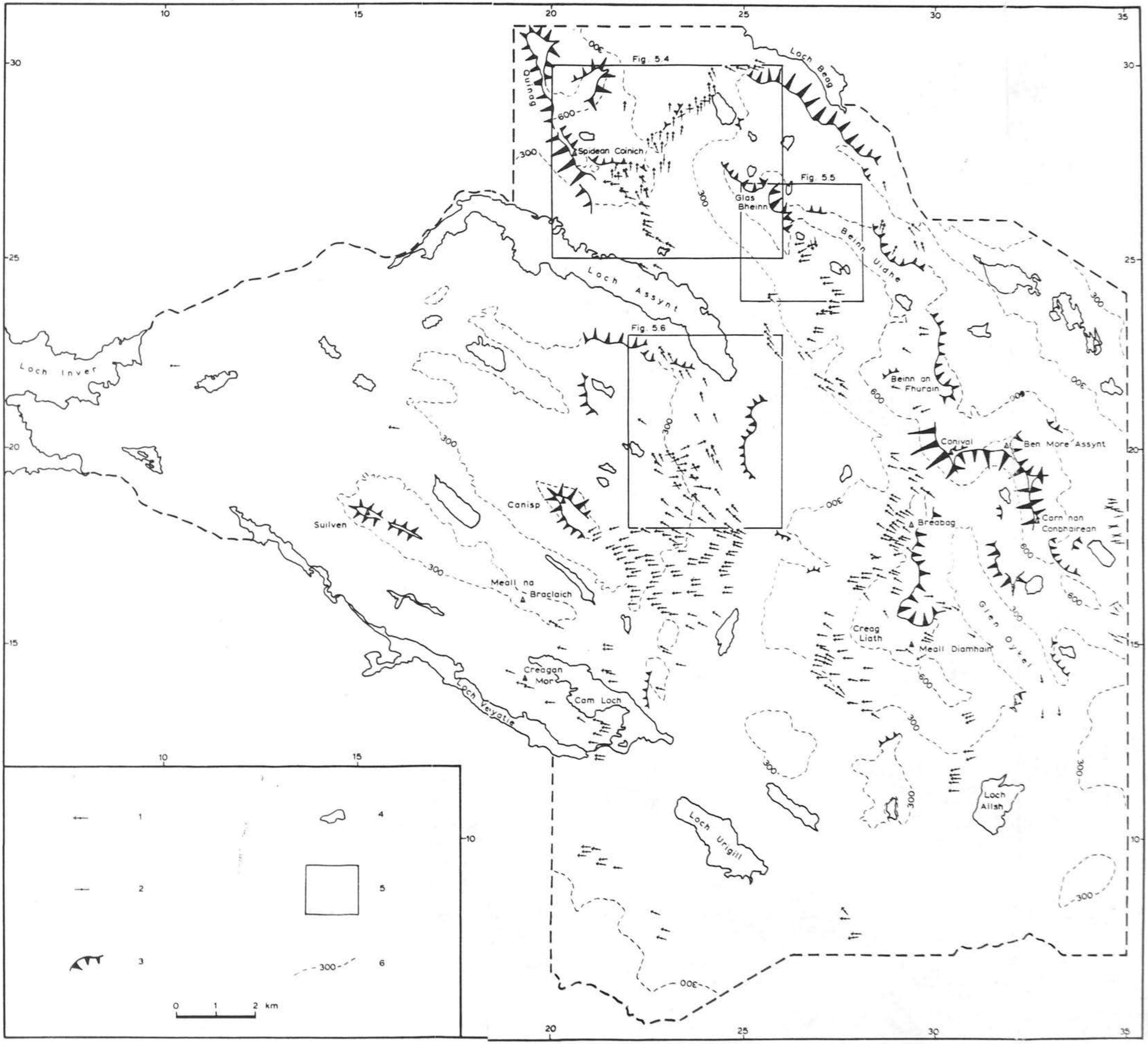


Fig. 5.4
Fig. 5.5
Fig. 5.6
Quinag
Spidean Coinich
Glas Bheinn
Loch Beag
Beinn Uidhe
Loch Assynt
Loch Inver
Beinn an Fhurain
Conival
Ben More Assynt
Breabag
Carn nan Conbhairean
Suilven
Canisp
Meall na Braclaich
Creag Liath
Meall Diamhain
Glen Oykel
Creagan Mor
Cam Loch
Loch Veyatie
Loch Urigill
Loch Ailsh
1
2
3
4
5
6
300
0 1 2 km

induce brittle fracture along lines of weakness, this having resulted in few striations on the higher ground where bedrock has been broken up by frost-shattering under periglacial conditions.

In addition to the quartzite areas, striae have been mapped on other rock types. At NC 15952052 and NC 10322227, striations were found on Lewisian gneiss which had recently been exhumed from beneath a thin till cover by rainwash. Shallow grooves were found on ice-moulded Lewisian gneiss at NC 28652689, to the north of Beinn Uidhe, and at three locations in Glen Oykel. Except for these isolated examples, no unequivocal glacial striations were found on Lewisian gneiss. Pseudo-striated surfaces of ice-moulded gneiss landforms are in fact due to postglacial etching-out of selective foliation planes. Several localities in the vicinity of Creagan Mor, between lochs Veyatie and Cam, possess a few shallow grooves on Torridonian sandstone, but postglacial weathering of the bedrock surfaces has removed all traces of any smaller subglacial abrasion features. Other shallow grooves on Torridonian sandstone were found at NC 21892896 and NC 23062887 in the Quinag area, and on the southern slopes of Meall na Braclaich, south-east of Suilven. No striations were found on Moine schists in the field area, although Read _et al_. (1926, p. 174) recorded numerous striae farther to the east, occurring "on thin strings or knots of vein-quartz, which generally project a quarter or at most half an inch beyond the present surface of the moutonnée schists in which they are found". Other lithologies in the area, such as the dolomites and 'Fucoid Beds', weather easily and therefore show no signs of glacial striae.

Other factors that influence the distribution of glacial striations in Assynt are the amount of exposure of glaciated bedrock, and the degree to which the area has been traversed in search of glaciated surfaces. A large proportion of the study area is covered in blanket bog and thin drift, which may be obscuring striated bedrock.

5.3.2 Glacial striae and former ice flow patterns in the Assynt area

Several ice flow trends are apparent in Fig. 5.3 : there is a main trend on which are superimposed a number of localised secondary trends. Across the bulk of the area, the pattern of former ice flow was westwards and north-westwards from the main mountainous area, down into the valleys and across spurs. The striae pattern shows that the ice was deflected around the higher ground (e.g. Quinag and Canisp), and that there was a strong convergence into the troughs between the hills, especially into the trough now occupied by Loch Assynt, and into the smaller valleys in the area between Canisp and Cul Mor, to the south-west of the study area. In the north, ice flowed to the north and north-west, and north of Beinn Uidhe striae point NNW and NNE. Striae of the narrow quartzite outcrop SE of Ben More Assynt trend approximately W - E, but no indication of ice-flow direction has been made in Fig. 5.3 as erratics in this area have moved both eastwards and westwards. In Glen Oykel, striae show that ice flowed from the major corries on its western side, and followed the main axis of the valley towards Loch Ailsh. In the Knockan basin area, the few striae noted on the thrust quartzite of the nappes along the Moine Thrust indicate former ice flow to the west and to the WNW.

Table 5.1 shows the highest altitudes at which striated bedrock has been found on some of the Assynt mountains. Although some of the locations mentioned occur within the area of periglacial frost-shattering, it is highly probable that the shattered quartzite blocks which mantle many of the summits in the area would once have

Highest occurrences of striated bedrock in the Assynt area

Mountain	Summit height	Highest striated surface	Grid ref.
Creag Liath	807 m	807 m	NC 28731585
Beinn an Fhurain	804 m	770 m	NC 28952166
Breabag	715 m	708 m	NC 29251800
Meall Diamhain	703 m	701 m	NC 29321485
Canisp	846 m	640 m	NC 21051795
Spidean Coinich	764 m	510 m	NC 21602122

Table 5.1

formed glaciated surfaces that would have yielded striations at even higher altitudes than those found in this study.

In three small areas, striations of a different direction from those of the main trend occur, sometimes on the same bedrock surface, and sometimes actually crossing those of the latter. In these locations, care was taken to attempt to determine the direction of ice flow and the relative ages of the two sets of striae from studies of their position on the rock outcrop, from local ice-moulding, from studies of erratics, and by trying to ascertain which set of striations was locally more dominant and therefore likely to be the more recent of the two.

(a) The Quinag area

On the quartzite dipslope of Spidean Coinich (Quinag), the main striae set shows divergence of ice around the mountain, pointing approximately westwards around the southern slopes above Loch Assynt, and northwards over the col between Quinag and Glas Bheinn. In several places on Druim na h-Uamha Moire and on the dipslope itself more striations can be found, indicating ice movement from west to east or east to west. It is argued that these secondary striae comprise two striation sets : if all these striae are taken as a single set it is difficult to explain ice movements in the area logically. The only erratics present are blocks of Lewisian gneiss and associated intrusives. The outcrops of these rocks lie to the north, east and south-east. Ice movement from the north is dismissed as very unlikely, as this region possesses no high ground which may have acted as a glacial source area. Ice movement from the south-east would seem to fit well with the main striation trend and also with the secondary NW - SE trend on the dipslope of Spidean Coinich.

Fig. 5.4 Former ice-flow patterns in the Quinag area.

(a) Glacial striae, as mapped.

(b), (c) and (d) Suggested sequence of former ice-flow directions in the area shown in (a).

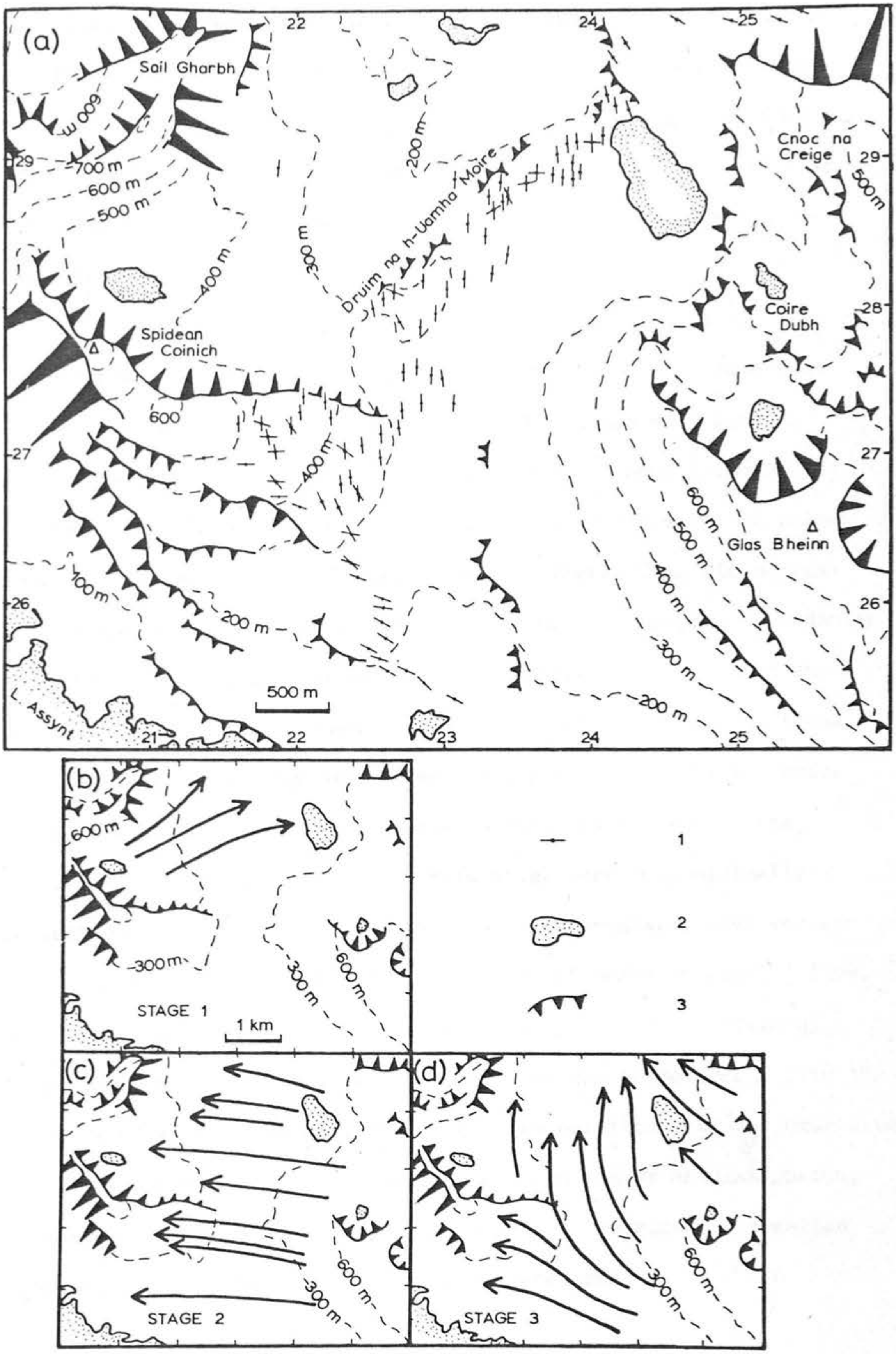
Key: 1. Glacial striae.

2. Major lochs.

3. Free faces.

Contours at 100 m intervals in (a) and 300 m intervals in (b), (c) and (d).

(diagram overleaf)



(a)
Sail Gharbh
Spidean Coinich
Druim na h-Uamha Moire
Cnoc na Creige
Coire Dubh
Glas Bheinn
L. Assynt
500 m
(b)
STAGE 1
1 km
(c)
STAGE 2
(d)
STAGE 3
1
2
3

The secondary striations on Druim na h-Uamha Moire could be seen as indicating ice flowing westwards, carrying Lewisian rocks from the Cnoc na Creige and Coire Dubh areas; however the striae would then seem to indicate ice flowing <u>into</u> the corrie beneath Spidean Coinich and Sail Gharbh. The more likely alternative is that the striae in this area indicate ice flowing <u>out</u> of the corrie and spreading over the Druim na h-Uamha Moire area.

The relative ages of the three striation sets must now be ascertained. The main trend diverging around the Quinag massif is doubtlessly the most recent of the three, being the most dominant and freshest striation set. Fig. 5.4 attempts to show graphically the most likely sequence of former ice flow directions in the Quinag area. In an early stage of glaciation, ice flowed from its source area in the corrie previously mentioned, probably carrying Torridonian erratics onto the quartzite and Lewisian outcrops to the north and north-east. At a later stage of glaciation, when the ice sheet had built up sufficiently to cover even the highest ground in the area, ice flow was more from the south-east. This was a powerful ice stream, as the striae relating to this stage were only minimally deflected by the presence of Quinag. At a later stage, when the ice was thinning, the topography had a greater influence on glacial flow, and the ice was deflected to the south and to the east of Quinag; this flow was the last one that affected the quartzite area. Both the ice flow of stages 2 and 3 carried Lewisian erratics onto the quartzite dipslope of Spidean Coinich from the southern slopes of Glas Bheinn, at the same time sweeping away all evidence of Torridonian erratics from stage 1 from the Druim na h-Uamha Moire area.

(b) The area to the south-east of Glas Bheinn

Just to the south and south-east of Lochan Bealach na h-Uidhe, dipping quartzite surfaces show two striation trends (Fig. 5.5). Striae following the main regional trend indicate ice movement from Beinn Uidhe, flowing SE to cross the spur to the south of Glas Bheinn. On certain of the rock surfaces facing north and north-west, which are largely protected from ice moving from the east and north-east, a further striation set can be seen trending down-valley. Nowhere is there good evidence from erratics or ice-moulding to suggest which of the striation sets is the more recent. Purely from the evidence of their lee-side position with respect to the ice flow that cut the striae conforming to the regional trend, it can be argued that the striae pointing down valley relate to an early phase of the same glaciation.

(c) The Loanan Valley

On the northward extension of the Canisp dipslope, on the western side of the Loanan valley opposite Stronchrubie, a series of quartzite surfaces shows very good examples of crossing striae (Fig. 5.6). The main regional trend in this area indicates a strong north-westward ice-flow. East of a line drawn from Meallan Liath Mor to Meallan Liath Beag, this set of striae is crossed by another associated with ice-moulding that indicates ice flowing north-eastward and eastward down the local slope. Within this area, the local striae set is dominant and the regional set tends to be well represented only in lee-side positions (i.e. on east- and north-east-facing slopes). At one location (NC 23371948), grooves conforming with the regional trend are cut by striae of the local set in the manner of Fig. 5.2, indicating that the latter is the more recent. The same picture is given by the

Fig. 5.5 Glacial striations mapped in the area to the south-east of Glas Bheinn.

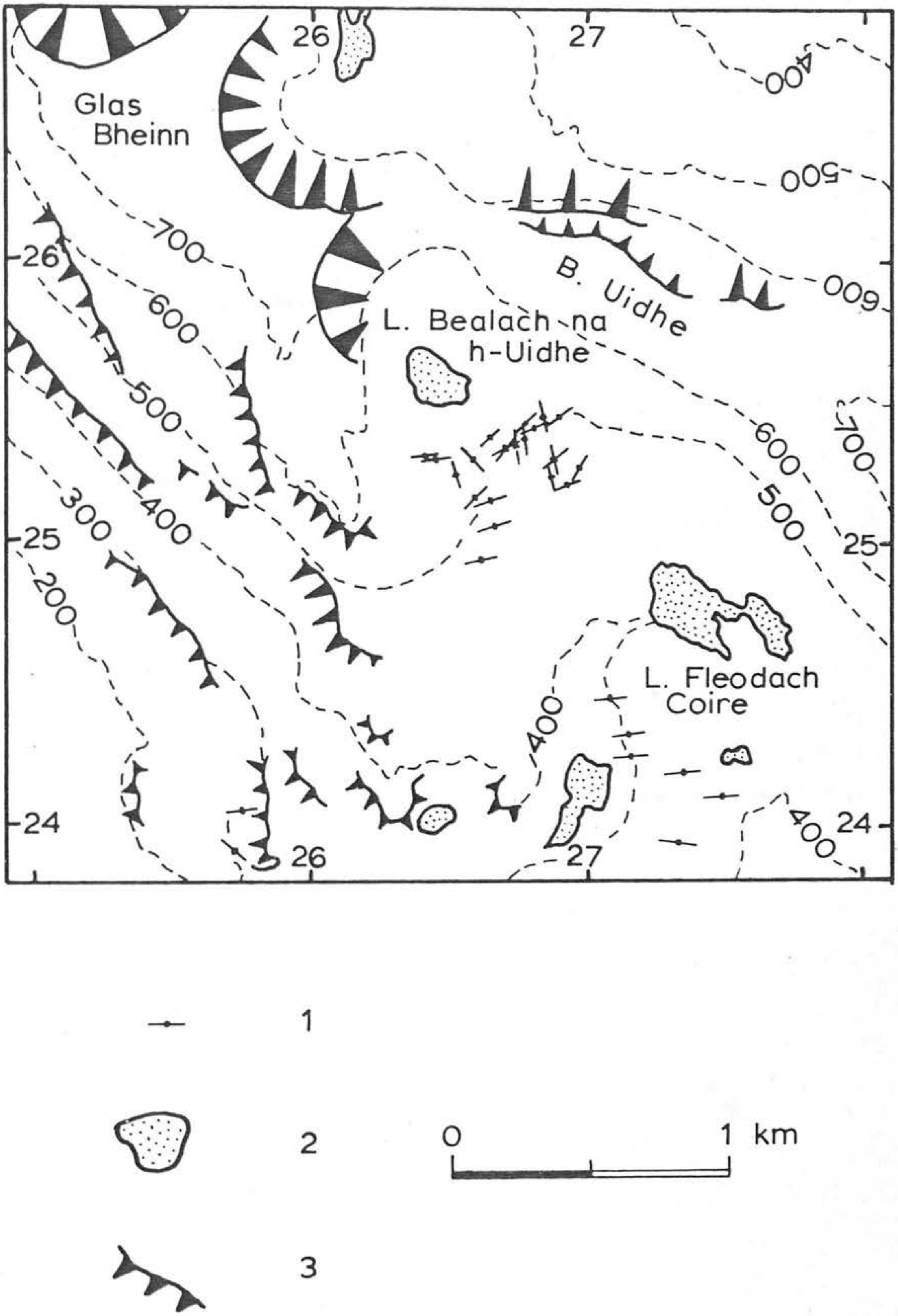
Key: 1. Striae trends.

2. Main lochs.

3. Free faces.

Contours at 100 m intervals.

(diagram overleaf)



Glas Bheinn
L. Bealach na h-Uidhe
B. Uidhe
L. Fleodach Coire
26
27
25
24
700
600
500
400
300
200
1
2
3
0
1 km

Fig. 5.6 Glacial striations mapped in the Loanan valley in the vicinity of Stronchrubie.

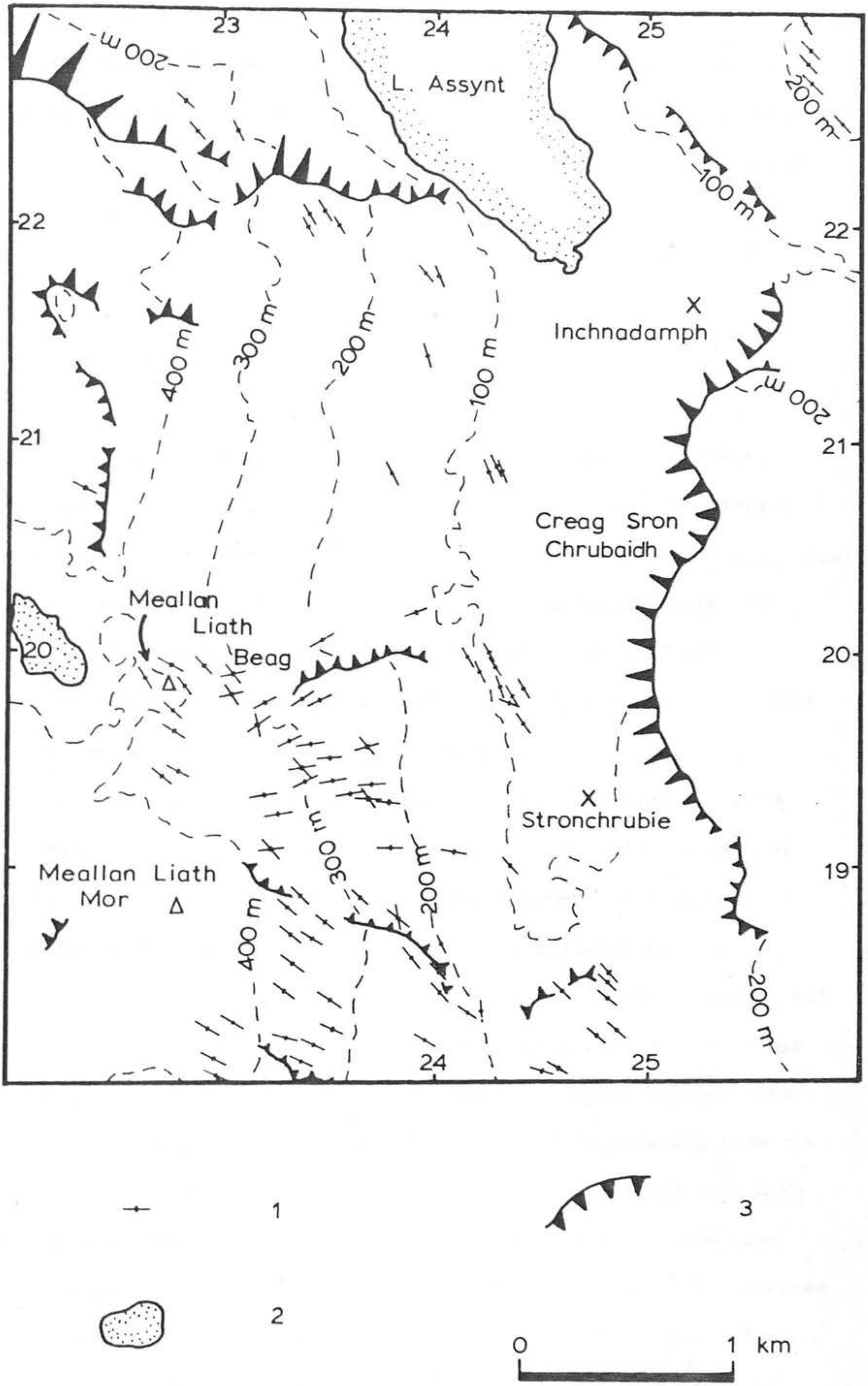
Key: 1. Striae trends.

2. Main lochs.

3. Free faces.

Contours at 100 m intervals.

(diagram overleaf)



23
24
25
22
21
20
19
200 m
L. Assynt
200 m
100 m
400 m
300 m
200 m
100 m
Inchnadamph
200 m
Creag Sron
Chrubaidh
Meallan
Liath
Beag
Stronchrubie
Meallan Liath
Mor
300 m
200 m
400 m
200 m
24
25
1
2
3
0
1 km

distribution of Canisp porphyry erratics (Fig. 4.2), which suggests that the downslope ice movement was the last to occur in this area. Section 6.3 discusses the above evidence, together with the distribution of other geomorphic features, to suggest that another small glacier occupied this area in the Lateglacial period.

5.4 CONCLUSIONS

It has been suggested above that the pattern of former ice flow in the Assynt area as shown by the orientation of glacial striae can be interpreted as indicating two phases of glaciation. The first phase saw ice covering the whole of the area, probably even the highest ground. Striated surfaces occur at over 800 m O.D., although weathering, especially frost-shattering processes, may have removed evidence of striation at higher altitudes. During this glacial phase, ice moved mainly northwards and westwards away from the main mountainous area. A few striated surfaces to the east of the mountains hint at a possible eastward movement, but further evidence is obscured by the lack of outcrops suitable for the preservation of striae. In the north, certain evidence suggests that the ice flow was more influenced by the topography at a later stage of the glaciation, probably caused by the thinning of the ice sheet.

A subsequent glacial phase, distinctly separate from the earlier one, caused small, localised glaciers to form and flow down the local slopes, thereby giving rise to other sets of striations which in places cross those of the earlier glaciation. This occurred on the west side of the Loanan valley opposite Stronchrubie, and in the corrie to the south-east of Glas Bheinn. As will be shown in section 6.3, other glaciers relating to this later period formed elsewhere in the study area.

The striae in the Assynt area indicate the presence of glacial ice whose basal layers actively slid across the glacier bed, armed with suitably hard abrasives; a certain amount of basal melting must have occurred to keep the abrasives in contact with the bed. Such conditions are found under temperate glaciers today, and it is therefore a reasonable assumption that the ice sheet represented by the striae of the earlier of the two phases was composed of temperate ice for at least part of the time it existed, and that the small glaciers of the last glacial phase were probably temperate throughout their existence. That conditions in the past were suitable for subglacial abrasion is at least in part due to the presence of quartzite, which dominates till lithology throughout the area (section 3.4), providing hard angular blocks which made good abrasives.

CHAPTER 6 : THE GLACIATION OF THE ASSYNT AREA

6.1 INTRODUCTION

The previous two chapters have discussed evidence for ice-flow patterns in the Assynt area. It is the aim of this chapter to discuss the conclusions reached in chapters 4 and 5 in the context of the present state of knowledge of the Scottish Quaternary. Therefore it is necessary to look at the work that has previously been done in the study area and its vicinity, to update such work in terms of current hypotheses, and to integrate the results of the present research.

6.2 PREVIOUS WORK

It is convenient to review the literature on the area in two sections, namely to deal with the results of work published prior to, and after, the widespread utilisation of radiocarbon dating. The introduction of this technique in 1949 enabled the construction of radiometric chronologies which allowed the results of work in one area to be directly compared with those in others. There is therefore a change in the type of work from that relying on relative chronologies to one based on radiometric dates, which occurs around 1965 in work on the NW Highlands.

6.2.1 Literature on the Assynt area prior to the development of radiocarbon chronologies

As early as 1831 MacCulloch noted that a large amount of erosion had occurred to carve out the Torridonian sandstone mountains of Sutherland and Ross-shire from the once continuous gritstone cover (MacCulloch 1831, p. 506), though he offered no explanation as to how this denudation had come about.

With the acceptance of the Glacial Theory in the decades following 1840, much evidence was accumulated as to the efficacy of glacial erosion in the development of highland landscapes in the British Isles. Due to the inaccessibility of the study area to early travellers, only Robert Chambers carried out any detailed work there. In a paper on the glaciation of parts of northern Britain (Chambers 1853) he dealt with specific examples from the Assynt area. He correctly attributed many of the geomorphic features to glacial processes, realising that many of them had been effected by a widespread glaciation of the whole area, whereas others seemed to be of more local origin. He believed, quite logically, that the more widespread glaciation was the earlier, but incorrectly interpreted the striae as indicating ice flowing from the west or north-west. He declared that the lack of erratic material brought westwards was due to the subsequent westward flow of local ice, transporting such rock fragments back to their source. The suggestion that the area was inundated by ice from the north and west is typical of contemporary ideas that the Great Ice Age was merely an equatorward expansion of the polar ice caps. However, Chambers appeared to contradict this view elsewhere in the paper when explaining the subsequent occupation of certain valleys by local ice. He wrote, "So much for the glaciation of this district, where there are bosoms amongst the hills and valleys running out of them, appropriate seats of local glaciers" (p. 249), seeming to indicate that he was aware that the mountains acted as glacial source areas rather than the polar regions. Despite such inconsistencies, Chambers noted many of the geomorphic features mentioned in the present research, such as the moraines in the Loanan valley (Chapter 3), and the largescale striation of the Cambrian quartzite surfaces (Chapter 5) which is paralleled by the long axis of Suilven and by many of the lakes of the Lewisian gneiss area (Chapter 3).

The officers of the Geological Survey were mapping in the area in the last two decades of the nineteenth century, and although they were mainly concerned with the solid geology, references to Quaternary features were made on the field maps and in the geological memoirs of many of the areas examined. The Assynt area is considered in the large memoir dealing with the whole of the NW Highlands (Peach et al. 1907), but this is concerned only with the solid geology. However, the Geological Survey officers did publish some information about the glacial history of the area. Peach & Horne (1892a) noted the westward transportation of Moine and Lewisian erratics onto Beinn an Fhurain and Breabag, concluding that the ice-shed of the last ice sheet was situated to the east of the present watershed. Two other articles by Peach & Horne (1892b, 1917) were mainly concerned with an excavation of one of the Creag nan Uamh bone caves (dealt with more fully in Chapter 10), but mention was made of the former glaciation of the area. In the 1917 paper it was noted that both striae and erratics indicated a westward movement of ice, and several examples of each type of evidence were given. Peach & Horne also noted the general lack of till cover in the area, and the apparent eastward carry of quartzite erratics into the Cassley valley, east of the Assynt mountains. They also stated that "there is evidence to prove that local ice streamed off the eastern slopes of Canisp and Beinn Gharbh" (Peach & Horne 1917, p. 333), which is doubtless a reference to the end moraine in the Loanan valley by Stronchrubie, and the crossing striae on the Canisp dipslope, although the actual evidence is not given.

Two Geological Survey memoirs (Read et al. 1926; Read 1931) dealt with the glaciation of the eastern part of the study area. A fourfold sequence of events was proposed as the basis for the interpretation of glacial features:

(a) a period of maximum glaciation, when the whole of northern Scotland was covered by an ice sheet whose ice divide lay to the east of the mountains;

(b) a period of waning glaciation with local centres of ice dispersion in the mountains;

(c) a period of independent valley glaciers;

(d) the retreat of the ice and a subsequent period of corrie glaciers.

Although more recent work has enlarged upon and changed this framework somewhat, it will be shown that much of this relative sequence of events is still valid. Specific mention was made in these memoirs of terminal and lateral moraines in Glen Oykel near Loch Ailsh, and several observations of erratics and striation trends reported in chapters 4 and 5 of this thesis were discussed.

Charlesworth (1955) identified numerous retreat stages across the Scottish Highlands as the ice withdrew after the last glacial maximum, but few of these glacial limits are supported by field evidence. Boyd (1956) suggested that, as the last ice sheet retreated, a large lake existed to the south of Suilven, including the areas presently occupied by lochs Fionn and Veyatie, draining to the sea by way of Gleann Sgoilte and Loch Bad na Muirichinn instead of by the present outlet, the River Kirkaig (Fig. 3.1). Unfortunately, much of Boyd's proposed sequence of events is unsubstantiated by field evidence and his arguments are based on too many unqualified assumptions. Godard (1965), looking at the whole of the NW Highlands, concerned himself mainly with the identification of erosion surfaces ("nivaux d'aplanissement") which he saw as dating from the Tertiary period. Godard assumed only minor modification of these surfaces under the subsequent ice sheets. In the course of his work in the Assynt area he mentioned the periglacial features on the mountain tops and the effects of the contrasting resistances to periglacial and glacial processes exhibited by Lewisian gneiss and Cambrian quartzite.

6.2.2 Synopsis of the Scottish Quaternary and past work in the Assynt area since the development of radiocarbon dating

Since the development of radiocarbon dating, there has been much progress in the understanding of the Scottish Quaternary. Reviews have been given by Sissons (1967, 1974a, 1976) and outlines are to be found in the introduction to the relevant INQUA Excursion Guides (Clapperton 1977; Price (ed.) 1977; Sissons 1977). Several more recent papers have further developed our understanding of the various phases of the last glaciation, including the reviews by Gray & Lowe (1977) and Sissons (1979b), and articles by Robinson & Ballantyne (1979) and Sissons (1981).

Until recently, consensual opinion was that rapid expansion of the last ice sheet occurred at c. 26,000 years B.P. until it reached its maximal extent at c. 18,000 years B.P., by which time it covered most (if not all) of Scotland and much of England, Wales and Ireland. Sissons (1981) has underlined the fact that the actual dating of the expansion of ice in Scotland is open to speculation, and little is known about the maximal extent of Scottish ice or whether ice existed in Scotland prior to 26,000 years ago. Evidence of the last (Ipswichian) interglacial and of the period of the last (Devensian) glaciation up to the glacial maximum is scanty due to the disruption caused by the ice advance. Views about certain readvances interrupting the decay of the last ice sheet from the glacial maximum have changed greatly in recent years. Postulated Aberdeen-Lammermuir and Perth Readvance stages (Sissons 1967) have been rejected (Paterson 1974; Sissons 1974a, 1976), but recent work in the NW Highlands has produced good evidence of an ice-marginal position, referred to as the Wester Ross Readvance (Robinson & Ballantyne 1979; Sissons & Dawson 1981). Abrupt drops in the marine limit at Stirling, in Loch Fyne and elsewhere near the head of the Firth of Clyde also suggest a minor ice-sheet readvance or stillstand (referred to in Sissons 1981). Beetle assemblages indicate a rapid climatic warming at c. 13,000 years B.P. (Bishop & Coope 1977), probably causing a swift decay of the remaining ice sheet. Radiocarbon dates on the oldest Lateglacial organic sediments at many sites across Scotland, including one of 12,810 $\pm$ 155 (Q-457) on the major watershed of the NW Highlands (Kirk & Godwin 1963), suggest that Scotland was largely ice-free by about 12,500 years B.P. (Sissons 1974a, 1976; Gray & Lowe 1977). Total deglaciation of Scotland prior to the Loch Lomond Stadial (c. 11,000 - 10,000 years B.P.) seems probable. A return to very severe climatic conditions during the Loch Lomond Stadial resulted in the formation of a large ice cap on the western Scottish Highlands, and glaciers in certain of the corries and valleys in other parts of the Highlands and elsewhere. Outside the glacial limits of this Loch Lomond Advance, periglacial conditions existed down to sea level. These arctic and sub-arctic environments eventually gave way to the present (Flandrian) interglacial - the Postglacial period - conventionally taken as starting at 10,000 years B.P.

Recent lithostratigraphic analysis of certain lochs in northern Scotland (Pennington et al. 1972; Pennington & Sackin 1975; Pennington 1975a, 1975b, 1977; Haworth 1976) included four lochs from the study area: Veyatie, Cam, Borralan and Ailsh. The first three showed the typical tripartite organic-inorganic-organic Lateglacial sedimentary sequence (i.e. they were not covered by glacier ice during the Loch Lomond Stadial). At Loch Ailsh, however, corers were unable to penetrate the thick minerogenic sediment attributable to a Loch Lomond Advance glacier that had crossed the area now occupied by the loch, only Postglacial sediments being sampled. Seven radiocarbon dates were obtained from the Cam Loch profile, although those of 12,787 ± 190 (SRR-251) and 12,436 ± 220 (SRR-250) were not in stratigraphic order and were considered to be contaminated (Pennington 1975a). A date of 12,956 ± 240 (SRR-253) relates to the development of a dwarf shrub tundra vegetation associated with the widespread climatic amelioration at c. 13,000 years B.P. However, deglaciation of the land surface, and its partial vegetation with pioneer plant species, was considered to have occurred well before 13,000 years B.P. (Pennington 1977). The nearest approach of birch woodland in the Lateglacial occurred in the Allerӧd chronozone, at which time the maximum pollen production of the ericaceous shrub vegetation occurred in the Assynt area. A sudden change from organic to minerogenic sedimentation, coincident with an equally sudden vegetation change to communities dominated by Artemisia and other taxa associated with disturbed soils, heralds the arrival of the more continental climate of the Loch Lomond Stadial, when periglacial processes disrupted stable soil conditions. Radiocarbon dates of 10,698 ± 490 (SRR-249) and 10,585 ± 450 (SRR-248) from this layer confirm this conclusion. Minerogenic sedimentation had ceased by about 10,400 years B.P. to be replaced by increasingly organic sediments associated with a rapid transition to closed plant communities, although the local spread of birch woodland did not take place until c. 9,000 years B.P. (Pennington 1977).

Sissons (1977a) mapped the glacial limits of the Loch

Lomond Advance in the field area, based on geomorphic evidence. He identified six former glaciers : north-east of Spidean Coinich on the Quinag ridge, to the north and to the north-east of Glas Bheinn, north of Beinn Uidhe, east of Beinn an Fhurain, and a large valley glacier in upper Glen Oykel.

Radiocarbon dates from a core taken near Loch Assynt by H.J.B. Birks and co-workers have recently been published (Switsur 1981). All the deposits were Postglacial in age with a basal date of 9,430 $\pm$ 150 (Q-1280). The initial expansion of birch woodlands is dated to 9,200 $\pm$ 120 (Q-1279), which compares favourably with the estimate of Pennington (1977).

6.2.3 Synthesis

As evidence prior to the last total inundation of the land surface by glacier ice is extremely sparse in Scotland as a whole, it is reasonable to assume that, excepting major elements of the topography (e.g. corries and U-shaped valleys), the majority of geomorphic features in the field area can be interpreted in terms of the last ice sheet, the Lateglacial Interstadial, the Loch Lomond Stadial and the present interglacial. Analyses of various geomorphic features in the Assynt area, detailed in the previous three chapters, have shown that indeed two distinctly separate glacial phases can be identified. A regional pattern of striae, erratics, ice-moulded landforms, etc., can be related to various stages in the existence of the last ice sheet. A less extensive glacial phase, superimposed on the above, can be attributed to the Loch Lomond Stadial, in view of the results of the work of Pennington and her co-workers.

6.3 THE GLACIERS OF THE LOCH LOMOND STADIAL

It has been suggested (Sissons 1980a, 1980b, 1981) that the Loch Lomond Advance can be seen as the correlative of an early stage in the build-up of a British ice sheet. This assumes that climatic conditions during the stadial were essentially the same as those at the start of a glaciation, and that such conditions prevailed from one glaciation to the next. It is therefore important to attempt to isolate the areas occupied by glacier ice in the Loch Lomond Stadial if one is to understand the initial build-up of the last ice sheet and its subsequent development.

Seven former Loch Lomond Advance glaciers were identified on the basis of geomorphic evidence, namely the presence of lateral and terminal moraines, the distribution of hummocky and fluted moraines, drift and boulder limits, the absence of fossil periglacial features which are assumed to date from this period, and the evidence of directional indicators such as striae and erratics.

Figs. 6.1 and 6.2 show the reconstructed Loch Lomond Advance glaciers, and the detailed evidence on which they are based.

(a) Quinag glacier (Fig. 6.1). The limits of this glacier have been largely interpolated between areas of bare glaciated bedrock surfaces, and those covered by abundant blocks of Torridonian sandstone and Cambrian quartzite. A small 500 m-long, discontinuous morainic ridge to the west of Druim na h-Uamha Moire is interpreted as a marginal feature. Although the distribution of hummocky and fluted moraines as mapped by the author varies somewhat from that shown by Sissons (1977a, Fig. 10, glacier 21), the glacial limits remain unchanged.

Fig. 6.1 Loch Lomond Advance glaciers in the northern part of the study area.

Key: 1. Free faces.

2. Areas of hummocky moraines.

3. Fluted moraines.

4. Striated bedrock.

5. Lateral and terminal moraines.

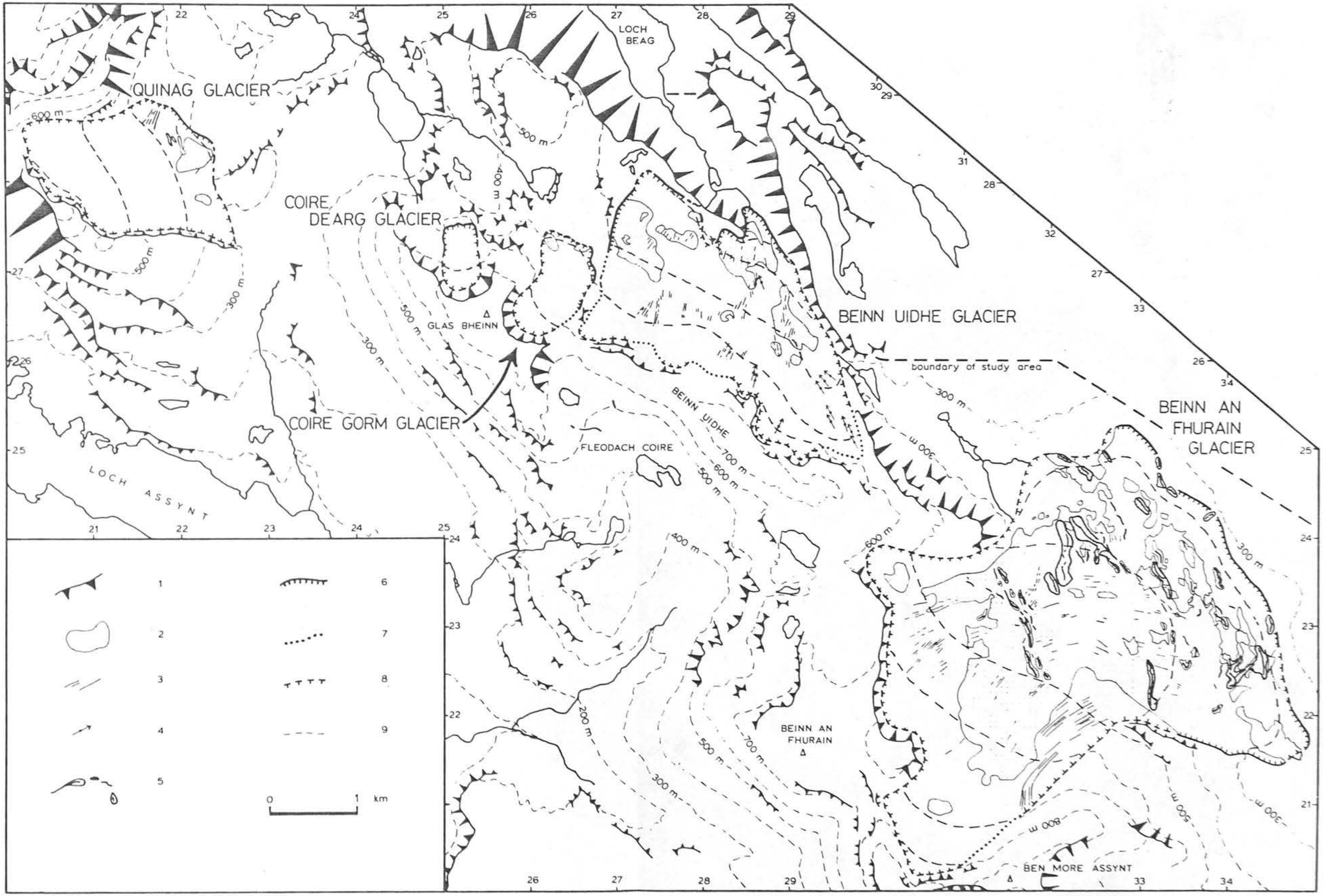
6. Limit of glacial drift.

7. Lower limit of periglacial features.

8. Interpolated glacial limit.

9. Contours at 100 m intervals (n.b. reconstructed glacier surface contours shown as a thicker dashed line).

(diagram overleaf)



QUINAG GLACIER
COIRE DEARG GLACIER
COIRE GORM GLACIER
BEINN UIDHE GLACIER
BEINN AN FHURAIN GLACIER
LOCH BEAG
GLAS BHEINN
BEINN UIDHE
FLEODACH COIRE
LOCH ASSYNT
BEINN AN FHURAIN
BEN MORE ASSYNT
boundary of study area
600 m
500 m
400 m
300 m
700 m
200 m
800 m
0
1 km
1
2
3
4
5
6
7
8
9

(b) Coire Dearg glacier (Fig. 6.1). The limit of this former glacier is well marked by the abrupt end to the spread of abundant quartzite debris that originated in the upper part of the corrie backwall. This boulder limit is remarkably clear, and hence no change is made here to the glacial limits mapped by Sissons (1977a, Fig. 4, glacier 26).

(c) Coire Gorm glacier (Fig. 6.1). A few discontinuous morainic ridges and hummocks help to define the northern limit of this former glacier, which is slightly larger than that mapped by Sissons (1977a, Fig. 4, glacier 27). The eastern and south-eastern limits are defined by the distribution of abundant quartzite blocks and frost-shattered bedrock surfaces. Other boundaries are interpolated.

(d) Beinn Uidhe glacier (Fig. 6.1). The source area of this former glacier is defined by rockwalls and periglacial features. The north-eastern margin is clearly marked by the distribution of frost-shattered rock surfaces, and by the drift limit in the Poll Amhluaidh area. A small discontinuous ridge above the Eas a' Chual Aluinn waterfall is interpreted as a marginal feature, and one can trace a clear drift margin, within which the ground is covered by hummocky and fluted moraine, into the valley of the Abhainn an Loch Bhig, where the limit abuts against a large roche moutonnée. Other margins are interpolated. The glacier so defined is larger and of different shape from the one mapped here by Sissons (1977a, Fig. 4, glacier 28); there are also marked differences in the distribution of hummocky and fluted moraines.

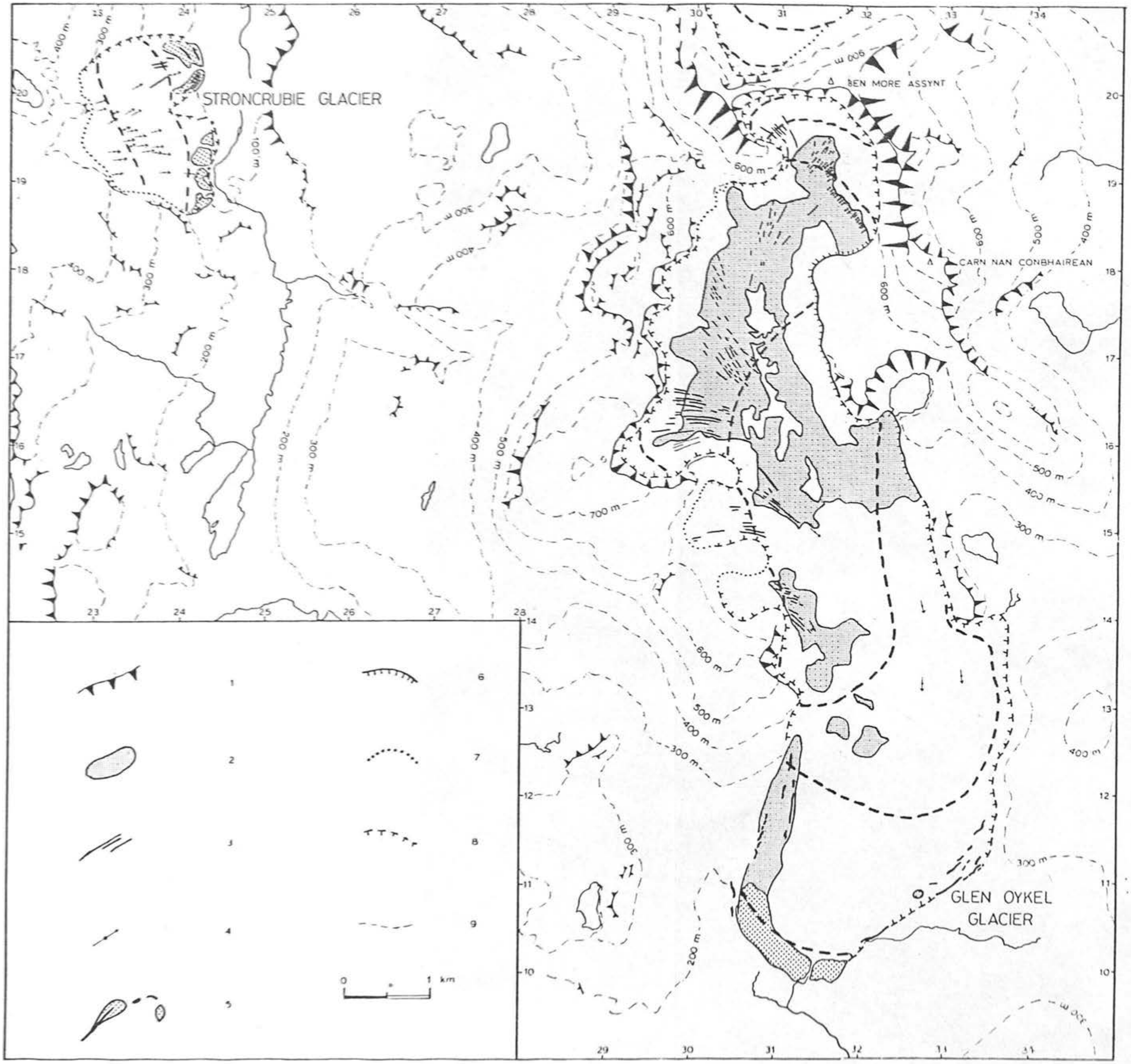
(e) Beinn an Fhurain glacier (Fig. 6.1). The northern limit of this former glacier has been drawn to include all the hummocky areas and morainic ridges in the vicinity of Gorm Loch Mor. The north-eastern and eastern margin follows the edge of the hummocky moraine and includes several short ridges interpreted as end moraines. A small part of the glacial limit here has been interpolated to fit the local contours. The southern limit is governed by the extent of periglacial features to the south-west, is interpolated to include the well-defined flutings of quartzite debris on the north side of Ben More Assynt, and follows the limit of hummocky moraines in the south-east. The glacial limit therefore closely follows that mapped in the area by Sissons (1977a, Fig. 4, glacier 29), except that the former glacier extends farther to the north-east by approximately 200 m. Within the eastern part of the area of hummocky moraines, several large ridges are prominent, and probably indicate ice-margin positions when stillstands or minor readvances interrupted the retreat of this glacier. The sinuous, branching, steep-sided ridge near Loch Bealach a' Mhadaidh was interpreted as an ice-cored moraine by Sissons (1977a), and probably marks the last stable position of the ice margin.

(f) Stronchrubie glacier (Fig. 6.2). A large end-moraine complex, consisting of three ridges spaced close together, clearly defines the north-western periphery of this former glacier, which Sissons (1977a) failed to map. Fluted moraines and striated surfaces occur within the area bounded by the end moraine. Although the other margins are not so clearly defined, the presence of crossing striae and erratics carried eastwards indicates that the former glacier also occupied the quartzite slopes south of the end moraine. The remainder of the eastern limit has been drawn along the easternmost extent of thick moundy drift, taking the distribution of ice-sheet striae into consideration; unfortunately the River Loanan has destroyed many of the marginal features here, and the former glacier may have extended

Fig. 6.2 Loch Lomond Advance glaciers in the southern part of the study area.

Key:
1. Free faces.
2. Areas of hummocky moraines.
3. Fluted moraines.
4. Striated bedrock.
5. Lateral and terminal moraines.
6. Limit of glacial drift.
7. Lower limit of periglacial features.
8. Interpolated glacial limit.
9. Contours at 100 m intervals (n.b. reconstructed glacier surface contours shown as a thicker dashed line).

(diagram overleaf)



STRONCRUBIE GLACIER
BEN MORE ASSYNT
CARN NAN CONBHAIREAN
GLEN OYKEL GLACIER
100 m
200 m
300 m
400 m
500 m
600 m
700 m
900 m
1
2
3
4
5
6
7
8
9
0
1 km

slightly farther eastwards. The western and southern limits are defined by the distribution of periglacial features and consideration of the presence or absence of Loch Lomond Advance striae.

(g) Glen Oykel glacier (Fig. 6.2). The limits of the large valley glacier that was mapped in Glen Oykel by Sissons (1977a, Fig. 5, glacier 30) are quite distinct, and no changes are proposed here after detailed work. Corries and the distribution of periglacial features restrict its western margin, and the well developed end moraine and lateral moraines define the southern boundary. Along the eastern side, the glacial limit follows a well defined drift limit or is interpolated.

There is evidence for local ice in a further two areas, although neither is attributed to the Loch Lomond Advance. Upper Fleodach Coire, on the south-east side of Glas Bheinn, possesses two small sections of a ridge composed of quartzite-rich debris (Fig. 6.1). This ridge is only c. 1 m high and, although it could be interpreted as a marginal morainic feature, it may also be seen as a drift-covered quartzite bedrock ridge since it closely follows the local geological structure. Within the area bounded by the ridge seven instances of glacial striae indicating a down-valley ice flow were noted, at variance with the widespread regional striation trend indicating ice flowing across the corrie from Beinn Uidhe towards Loch Assynt (Fig. 5.5). These local striations were all found in lee-side positions with respect to the ice sheet flow (section 5.3.2(b)), and hence may relate to an early phase of the same glaciation. This, together with the equivocal nature of the ridge of quartzite debris, leaves the suggestion that a Loch Lomond Advance glacier occupied this area open to some doubt, and hence no such glacier is shown in Fig. 6.1.

The large area of moundy till to the east of Ben More Assynt and Carn nan Conbhairean (Fig. 3.3) has been described in section 3.4.2(c). The mounds do not exhibit the steep conical appearance of hummocky moraines, and have therefore not been ascribed to a Loch Lomond Advance glacier. The presence of meltwater channels and other marginal features suggests an ice limit near Duchally (NC 388170) in the Cassley valley to the east of the study area (J.B. Sissons, pers. comm.), which further work may show to relate to the distribution of the till mounds, indicating the presence of a former glacier nurtured in the corries on the east side of the above mountains. At present, however, the age of such a glacier is unknown.

6.4 THE LAST ICE SHEET IN THE ASSYNT AREA

Having identified the areas occupied by glacier ice during the Loch Lomond Stadial, all glacial evidence outside these areas is assumed to relate to the last ice sheet. From a careful consideration of the geomorphic evidence from the Assynt area, together with the results of work from neighbouring areas, it is possible to determine much useful information about that portion of the last ice sheet that covered this part of northern Scotland. Of paramount importance are the way in which the ice sheet developed, the extent and thickness of the ice sheet at the glacial maximum, its surface form and thermal regime, and the way it retreated.

6.4.1 Ice sheet development

It has been shown that Loch Lomond Advance glaciers developed on the northern and eastern sides of the Assynt mountains, and in the Loanan valley. Taking the Loch Lomond Advance as a model, it is likely that a similar pattern developed at an early stage of the last ice sheet. Corries that did not act as source areas for Loch Lomond Advance glaciers may, however, have done so in this earlier phase of glaciation. Initial expansion of glacier ice would therefore have been northwards and eastwards onto the Moine schist area from the mountains, southwards down Glen Oykel towards the Cromalt Hills, and into the Loanan valley (Figs. 6.3(a) and (b)). Such an initial ice-flow pattern is supported by the southward dispersal of Lewisian erratics down Glen Oykel and by the carry of Lewisian and quartzite erratics onto areas east of the Moine Thrust, described in chapter 4. Discrete glaciers would have merged as they continued to expand. With the thickening of the ice, a stage would have been reached when it overtopped the mountain ridge, and started to flow both east and west from an ice divide situated to the east of the mountains (Figs. 6.3(c) and (d)). At the glacial maximum, ice flowed generally to the WNW in the southern part of the study area, carrying blocks of Moine schist to the west of the Moine Thrust. The lack of Moine erratics in the north of the area is explained by ice flowing in a more north-westward direction. At all times the ice divide was situated to the east of the mountains, over the Moine schist area.

6.4.2 Ice extent and thickness at the glacial maximum

The presence of quartzite blocks, with surfaces exhibiting striations, friction cracks and ice-moulded features, amongst the shattered debris on the ridge between Ben More Assynt and Conival indicates that the highest parts of the field area have at some stage been ice-covered. If it is reasoned that the blockfields relate to

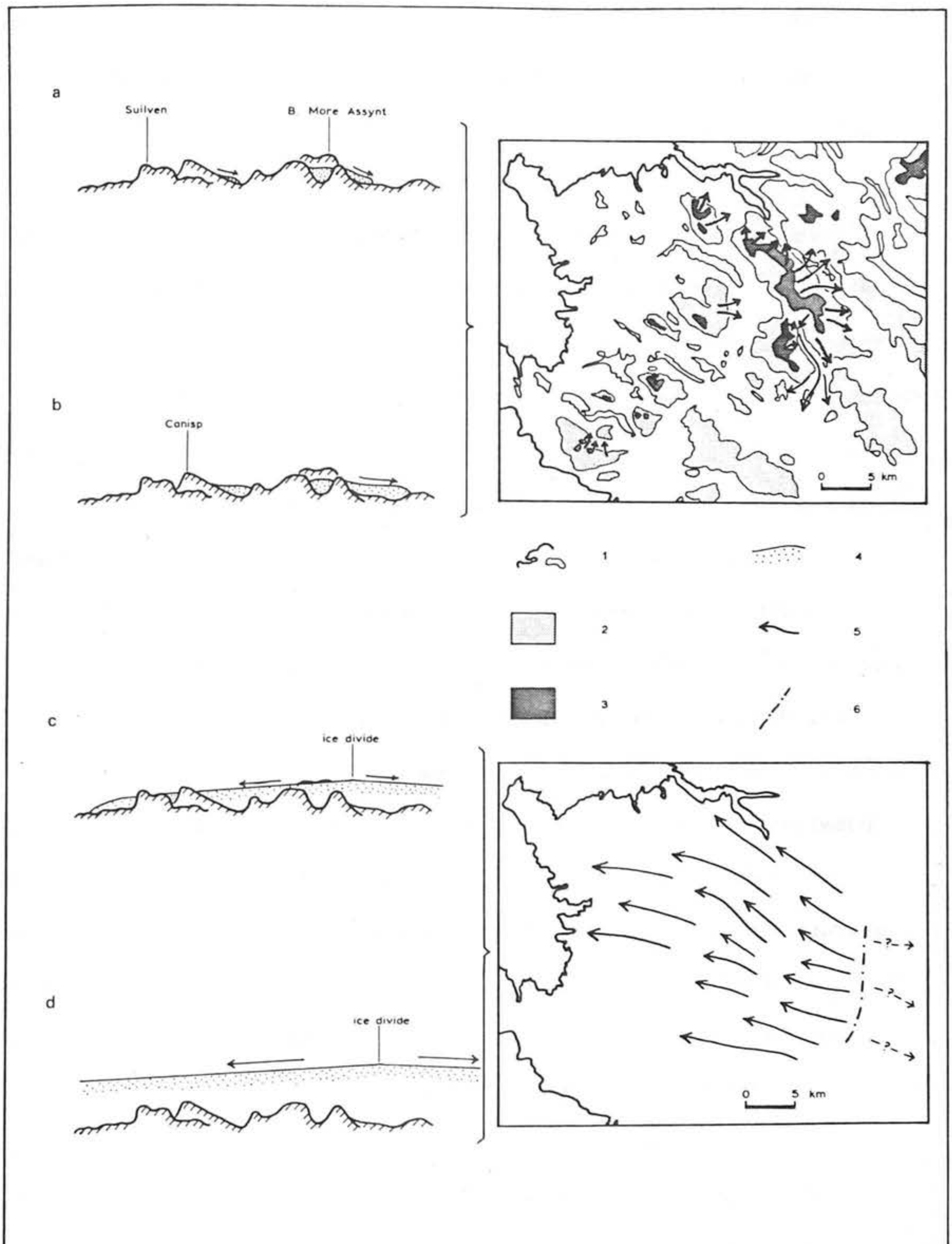
Fig. 6.3 Model of the development of the last ice sheet in the Assynt area.

(a) and (b): initial build-up of glaciers in the glacier source areas.

(c) and (d): continued ice build-up leads to the development of an ice divide when ice overtops the mountain ridge and flows both eastwards and westwards.

Key:
1. Coastline and main lakes.
2. Land 300-600 m O.D.
3. Land over 600 m O.D.
4. Glacier ice.
5. Direction of ice flow.
6. Ice divide.

(diagram overleaf)



a
Suilven
B. More Assynt
b
Canisp
c
ice divide
d
ice divide
0 5 km
1
2
3
4
5
6
0 5 km

the intense periglacial conditions of the Loch Lomond Stadial (section 3.5.2), then the original ice-moulding of the above-mentioned blocks must pre-date this event. As there is no evidence to the contrary, it can be assumed that complete glacial inundation of the area was achieved during the existence of the Late Devensian ice sheet. Thus a minimal altitude for the surface of the last ice sheet at its greatest extent must be considered to have been of the order of 1100 m O.D. (ignoring the effect of isostatic depression of the land). The distribution of glacial erratics and ice-roughened terrain indicates that the last ice sheet extended at least to the west coast, a distance of at least 27 km along the glacial flow lines from Ben More Assynt.

These minimal values are likely to be gross underestimates, as work in other areas has shown that mainland erratics have been found on North Rona (70 km north-west of Cape Wrath) and Sula Sgeir (20 km farther west), and erratics from the east exist on the Flannan Islands (30 km west of the Outer Hebrides), but that mainland ice failed to reach St. Kilda (70 km west of the Outer Hebrides) (Sissons 1976, p. 73). Flinn (1978) contended that evidence for an eastward flow of ice on the east side of the Outer Hebrides precluded the covering of those islands by the last mainland ice sheet, relegating the presence of mainland erratics on the west coast of North and South Uist to a glacial inundation from the mainland during a previous glaciation. However, Sissons (1980b) has argued that Flinn's explanation is unsatisfactory and has suggested that ice from the mainland did, in fact, cross the Outer Hebrides some time in the Devensian, with separation from the mainland ice during ice decay allowing the Hebridean ice to flow east and west from an ice divide that became established along the west side of the island chain.

Several attempts have been made to compare the profiles of modern ice sheets with theoretical profiles based on various ice sheet models (Paterson 1969, pp. 145-160). Nye (1952) suggested that inland from a known margin, the height of any point on the ice surface can be found from the formula

$$h = (2h_o s)^{1/2}$$

where h = the ice surface altitude, h_o = a constant of 11, and s = the horizontal distance from the margin (all in metres). Compared with other theoretical profiles and real profiles, Nye's parabolic curve slightly overestimates the slope near the centre of an ice sheet, and therefore slightly overestimates the altitude of the ice divide (Sugden & John 1976, Fig. 4.4). Nevertheless, one can utilise Nye's model to arrive at order-of-magnitude estimates for the form of the last ice sheet in the vicinity of the study area. Indeed, in view of the present lack of detailed information about the extent and thickness of the Scottish ice sheet, more complicated, supposedly more accurate, mathematical solutions to the present problem need not be considered.

Taking the minimal ice-surface altitude (h) above Ben More Assynt as 1100 m above present sea level (see previous discussion), the minimum distance (s) to the ice margin using Nye's equation is 55 km, which places the margin approximately half-way between the west coast of Sutherland and the east coast of Lewis. If one considers the other extreme, namely that the last ice sheet reached the Flannan Islands when it was at its maximum, it is possible to calculate a maximal ice-surface altitude at the ice divide. Assuming an ice divide about 10 km east of Ben More Assynt at the ice maximum, the distance (s) to an ice margin just west of the Flannan Islands is 200 km. Using this value, Nye's equation gives the altitude of the ice sheet surface at the ice divide as 2100 m above present sea level. Hence, using Nye's model and geomorphic evidence, values for the extent of the last ice sheet vary from a minimum of 55 km west of Ben More Assynt to a maximum somewhere between the Flannan Islands and St. Kilda, and estimates of heights for the ice sheet surface at the ice divide vary from 1100 m to 2100 m above present sea level. This model however assumes a horizontal base, an ice sheet in a steady state deforming as a perfect plastic material, with a constant ice temperature and accumulation spread evenly over its surface. No attempt has been made to allow for glacio-isostatic depression of the land surface and lower relative sea levels during the glaciation, and no account taken of the effect of the ice floating as it crossed the present area of the North Minch and the sea to the west of the Outer Hebrides.

Sissons (1981) has advised caution when considering the maximal extent of the last ice sheet : as he pointed out, there is no conclusive evidence that any of the westward-transported erratics on the Flannan Islands, Sula Sgeir or on the Outer Hebrides were carried there by the last ice sheet.

6.4.3 Direction of flow and surface form

Analysis of glacial striations and the distribution of glacial erratics in the study area has enabled a fairly detailed picture of the glacial flow lines of the last ice sheet to be constructed (Fig. 6.4(a)). The most detailed evidence is given by glacial striae, which show variations in flow paths in certain localities. Due to the essentially transitory nature of striae, the fullest striation pattern probably relates to a late stage in the existence of the last ice sheet, when the ice was thinning and topographic irregularities were influencing the glacial flow paths. However, evidence of linearly-eroded bedrock forms and the vestiges of striation trends relating to earlier phases in the existence of

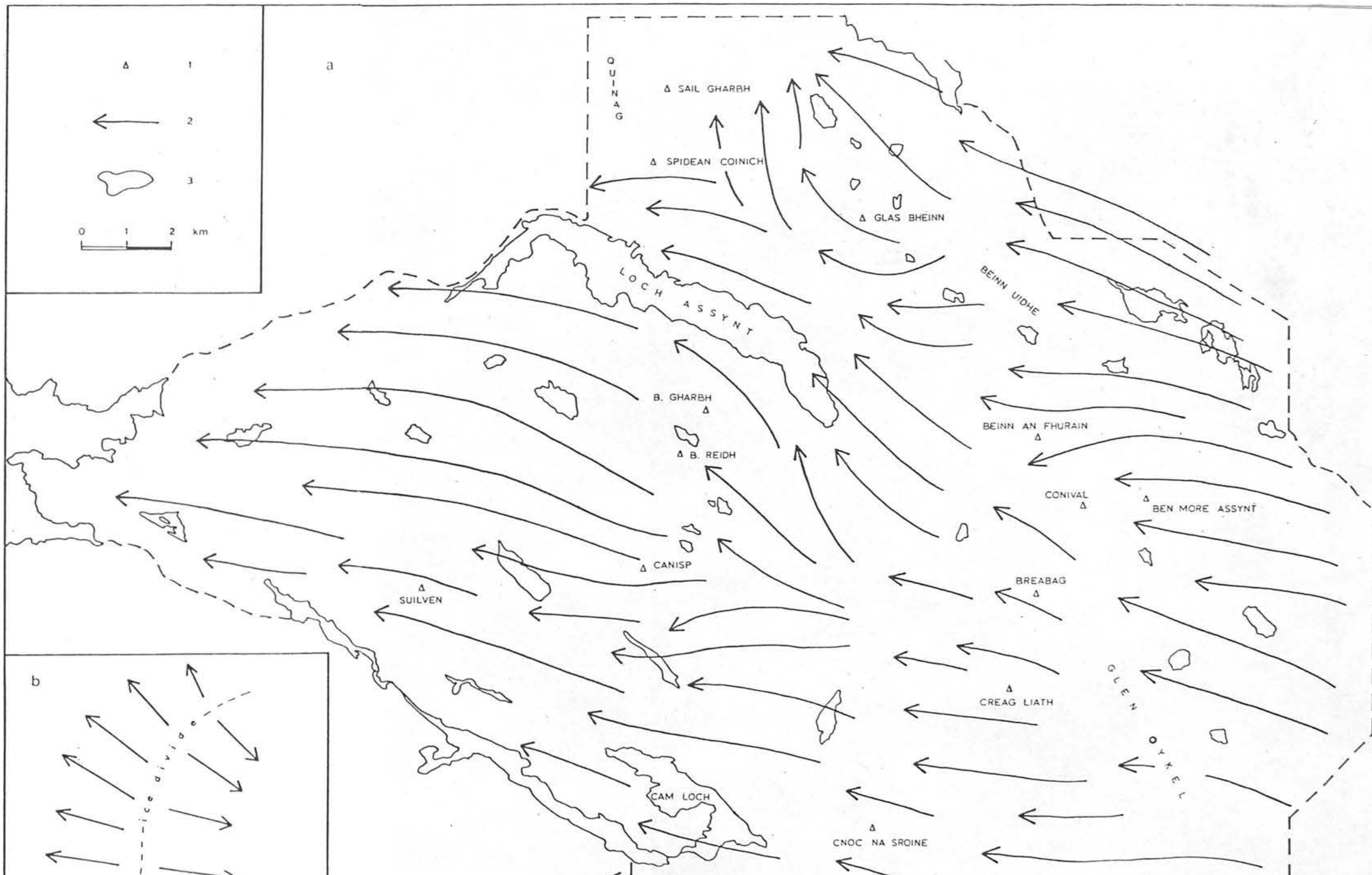
Fig. 6.4 (a) Patterns of ice flow of the last ice sheet over the Assynt area.

Key:

1. Major summits.
2. Lines of ice flow.
3. Present location of the coastline and major lochs.

(b) and (c) Suggested configurations of the ice sheet surface that would have resulted in such an ice-flow pattern over the Assynt area.

(diagram overleaf)



1
2
3
0 1 2 km
a
QUINAG
SAIL GHARBH
SPIDEAN COINICH
GLAS BHEINN
LOCH ASSYNT
BEINN UIDHE
B. GHARBH
B. REIDH
BEINN AN FHURAIN
CONIVAL
BEN MORE ASSYNT
CANISP
SUILVEN
BREABAG
GLEN OYKEL
CREAG LIATH
CAM LOCH
CNOC NA SROINE
AN STUC
CROMALT HILLS
b
ice divide
c
DOME
study area

the ice sheet, point towards a general flow of ice across most of the study area in a WNW and north-westward direction, largely uninfluenced by the underlying topography, although some divergence probably occurred around the highest mountains.

The basal flow under a thick ice sheet reflects its surface form, with ice tending to flow perpendicular to the ice-surface contours and away from the ice divide. The ice-flow pattern shown in Fig. 6.4(a) can be explained by two possible configurations of the ice sheet surface. Firstly, if the study area was covered by an ice sheet whose ice divide trended approximately SSW - NNE and curved away to follow a more SW - NE trend north of about northing 22 (Fig. 6.4 (b)), such a flow pattern would ensue. Secondly, the study area may have lain across the north-western portion of an elongated dome of ice whose highest point was somewhere to the south (Fig. 6.4(b)). Each situation is equally plausible on the evidence of striae and erratics presented in this study. Only further work in areas to the north and east will determine which of these situations applied.

6.4.4 Thermal regime

The only direct evidence pertaining to determination of the thermal regime of the last ice sheet in the field area is provided by the widespread presence of glacial striae, indicating basal sliding at the time of their formation. Basal sliding is normally associated with temperate ice, with basal temperatures at or above the pressure melting point.

Recently two attempts have been made to model the last ice sheet: Boulton *et al*. (1977) attempted to reconstruct the whole of the British Late Devensian ice sheet, and Gordon (1979) tried to model a section through the last ice sheet from Skye to the Black Isle. In view of the limited information about the extent of the last ice sheet, ice thicknesses and the relationship to the Scandinavian ice sheet, together with little or no information about ice-surface temperatures, and net accumulation distribution and rates, the model of Boulton _et al_. (1977) seems a trifle premature and their conclusions should be interpreted with some scepticism. Indeed the authors themselves admit that their model predicts unacceptably high ablation rates to the west of Scotland, in the North Sea and in the Irish Sea, and cannot account for the lobe of ice that deposited till over the coastal areas of eastern England as far south as Norfolk. The glacial flow lines over the Assynt area shown in Fig. 17.2 of Boulton _et al_. (1977, p. 234) do not accord with those indicated by the evidence presented in chapters 4 and 5 of this thesis. Their model showed that ice attained 1800 m over the Scottish Highlands and was predicted to be cold-based over much of its area, but temperate near the margins. Gordon (1979) also predicted a large zone of basal freezing for the central area of the ice sheet over northern Scotland. Though Gordon made numerous assumptions, his calculations for various values of ice thickness, the geothermal heat flux and accumulation rates suggested that these do not greatly affect this pattern of basal temperature conditions, but "errors in the estimates of mean annual temperature at the ice margin and ice-surface temperature lapse-rate will have much more significant effects and, depending on their magnitude and direction, basal thermal conditions might range from completely freezing to completely melting over the area considered" (Gordon 1979, p. 338).

Both ice-sheet models predict that a wide zone on either side of the ice divide would be one of net basal freezing. Such a situation is to be expected under polar ice sheets (Hooke 1977). The presence of glacial striae in the Assynt area within 10 km of the postulated ice divide, indicating net basal melting, is therefore problematic: either the predicted thermal conditions at the base of the last ice sheet at its maximal extent are wrong, or the striae mapped in the field area do not relate to the glacial maximum but to an earlier or a later phase of the ice sheet when temperate ice covered the area. It has already been suggested in section 6.4.3 that the striae in the field area relate to a period after the glacial maximum when the ice was thinning and the underlying relief exerted some control on ice-flow patterns. At this time, the Assynt hills would still have been buried beneath a considerable thickness of ice and an ice-shed zone would still have existed close to the mountains. Hooke (1977) has suggested that an increase in ice-surface temperatures or a drop in the accumulation rate, with subsequent thinning of an ice sheet and changes in the velocity field, would tend to raise basal ice temperatures. Therefore, if it could be shown that a significant change of either sort occurred after the glacial maximum, this might explain why temperate basal ice existed in close proximity to the ice divide at this time.

6.4.5 Deglaciation

Several recent papers have been concerned with the mechanism by which deglaciation from full glacial conditions occurred in the Late Devensian, based on the evidence from North Atlantic deep-sea cores (Ruddiman & McIntyre, 1981a, 1981b, 1981c). During the interval c. 20,000 - 13,000 years B.P., the oceanic polar front was located off northern Spain (at 40° - 45° N). The period 20,000 - 16,000 years B.P. is seen as the glacial maximum. Between 16,000 and 13,000 years B.P., Ruddiman and McIntyre suggest that the oceanic record indicates a rapid volumetric deglaciation, being the period of maximum influx of meltwater and icebergs from the ice sheets of the Northern Hemisphere. During this period, predicted summer insolation values were slowly rising, but pollen and beetle evidence indicate the presence of cold, dry air masses until _c_. 13,000 years B.P. Ruddiman and McIntyre explain this early thinning of the ice sheets by increased calving due to rising sea levels, and they attribute starvation of the ice sheets to reduced moisture extraction by the atmosphere as a low-salinity surface layer of cold water formed, probably freezing over in winter. The marked climatic amelioration noted at _c_. 13,000 years B.P. (Coope & Brophy 1972; Coope 1977) was associated with the replacement of polar water with warmer water in the eastern Atlantic adjacent to the British Isles; the position of the polar front at this time was far to the north-west (Ruddiman & McIntyre 1973; Ruddiman _et al_. 1977). However, by 13,000 years B.P. much of Scotland had been deglaciated and hence the cause of the ice retreat must be assigned to the suppression of nourishment in the form of snow. The striae near to the ice-shed zone in the Assynt area (section 6.4.4) could relate to the subsequent thinning of the ice sheet.

The last glacial retreat was in part an active one, as shown for example by the raised beaches associated with sequential ice-marginal positions in south-east Scotland. However as widespread downwastage of the last ice sheet continued, ice in many areas became stagnant and decayed _in situ_. In the Assynt area there is little geomorphic evidence relating to this deglacial period. No evidence pertaining to the Wester Ross stage identified farther south was found. Deglaciation of the region occurred prior to 13,000 years B.P. as indicated by the dated biostratigraphy of the Cam Loch core.

CHAPTER 7: PALAEOCLIMATE OF THE LOCH LOMOND STADIAL IN THE ASSYNT AREA

7.1 INTRODUCTION

Many of the glacial limits of the Loch Lomond Advance in Scotland and the English Lake District have been mapped by Sissons and co-workers (references given in Sissons 1980a; Cornish 1981). Attempts to make palaeoclimatic inferences from the reconstructed glaciers have been made to varying degrees of refinement (e.g. Sissons 1974b, 1977a, 1977b, 1979a, 1979c, 1980a; Sissons & Sutherland 1976; Ballantyne & Wain-Hobson 1980; Cornish 1981).

The total evidence has recently been summarised by Sissons (1979c, 1980a). He inferred that snowfall during the stadial was associated mainly with southerly and south-easterly winds, probably preceding warm and occluded fronts approaching from the south-west. However south-westerly winds were, on the whole, more common, playing a large part in redistributing snow from the surrounding high ground onto glacier surfaces. Precipitation was high in the western Grampians and southern part of the NW Highlands, but relatively low in the remainder of the latter area and very low in Speyside and the NW Cairngorms. This pattern was largely attributed to cyclones following more southerly tracks than at present, in turn related to the more southerly position of the oceanic polar front during the stadial, as well as to the likely presence of blocking anticyclones over Scandinavia, the freezing of the North Sea in winter and the orographic effect of the large ice mass that formed in the W Highlands. Sissons (1980a, p. 42) was careful to point out that palaeoclimatic inferences from the distribution of Loch Lomond Advance glaciers must be qualified in the knowledge that the oceanic polar front was not stationary during the stadial, and hence only "inferences about average conditions over a period of varying duration" can be made.

Sissons (1977a, Fig. 11) included a linear trend surface of the regional firn line, based on his mapping of the Loch Lomond Advance limits in northern Scotland. However, he did not present detailed climatic inferences of the kind made elsewhere (e.g. Sissons & Sutherland 1976). Following the present remapping of the former glaciers in the Assynt area (section 6.3), equilibrium firn lines were recalculated and attempts made to explain the evident spatial variation.

7.2 METHODOLOGY

As a first step the Loch Lomond Advance glaciers in the present study were reconstructed. Postulated glacier surfaces were contoured at 50 m intervals. (Only contours every 100 m are shown in Figs. 6.2 and 6.3.) Contours were drawn following the guidelines suggested by Sissons (1974b, 1977a), to approximate normal contour curvature on modern glaciers and to take account of the field evidence. Contours were, for example, drawn as near as possible perpendicular to the orientation of fluted moraines, which are likely to represent the latest direction of ice movement in these areas. Once glacier surfaces had been contoured, equilibrium firn lines were calculated using the equation given in Sissons (1974b), with areas between contours being measured using a grid of squares with sides equivalent to 100 m.

Once equilibrium firn lines had been calculated, an attempt was made to quantify the various factors affecting their altitude and the location of the glaciers. Potential avalanching areas were delimited, being defined as slopes of 20° or greater leading directly down to the glacier surface above the equilibrium firn line (after Sissons & Sutherland 1976). Potential snow-blowing areas were arbitrarily defined as those which "include all points that could be connected to the glacier surface with a continuous downhill slope along the ground surface" (Sissons & Sutherland 1976, p. 330). Snow-blowing areas were measured for winds from each of the four quadrants. The potential influence of direct insolation was calculated using the method of Sissons & Sutherland (1976). The insolation factor thus arrived at cannot give an absolute value of direct insolation absorbed by the glacier surface as the method assumes clear-sky conditions. Nevertheless the relative importance of direct insolation in the mass balances of each former glacier can be discussed from the values obtained (Sissons & Sutherland 1976; Sissons 1977b).

7.3 RESULTS

The results of these analyses are presented in Table 7.1. The mean equilibrium firn line altitude for the area is 446 m, though calculated values range from 239 m to 572 m. Such a large spread of values is mainly due to the equilibrium firn line of the Stronchrubie glacier which is considerably lower than the others. The altitudinal variation of firn lines for the former Assynt glaciers must be explained with respect to aspect, amount of precipitation and wind direction. Although his calculated regional firn line for the whole of northern Scotland increased altitudinally from NW to SE, Sissons (1977a, 1980a) noted that within individual mountain groups there was

GLACIER	GLACIER AREA (km^2)	EQ. FIRN LINE (m)	INSOL. FACTOR	AVAL. AREA (km^2)	SNOW-BLOWING AREAS (km^2)			
					SW	SE	NW	NE
QUINAG	2.608	446	11.9	0.557	0.276	0.060	0.246	0.222
COIRE DEARG	0.301	572	9.7	0.137	0.104	0.190	0.030	0.034
COIRE GORM	0.690	561	10.3	0.173	0.290	0.090	0.175	0.025
BEINN UIDHE	5.013	423	10.2	0.568	1.310	1.198	0.953	0.078
BEINN AN FHURAIN	13.640	459	13.5	1.013	2.858	1.622	1.047	0.685
STRONCHRUBIE	2.360	239	14.4	0.105	2.510	0.350	1.660	0.320
GLEN OYKEL	23.010	420	14.5	1.577	2.807	2.067	2.774	1.812

TABLE 7.1

Palaeoclimatic data

a tendency for the local firn line to rise from south to north. Such a pattern is clearly apparent in the Assynt area (Table 7.1), where firn lines increase in altitude from the SSW towards the NNE.

Calculated insolation factors vary quite considerably over the study area (mean = 12.1 $\pm$ 1.9). The Beinn an Fhurain, Stronchrubie and Glen Oykel glaciers all have relatively high insolation factors, indicating that they were fairly unfavourably located in terms of aspect.

It is difficult to use the values presented in Table 7.1 to compare the different areas which potentially contributed to the accumulation of snow on the glaciers, as glacier sizes varied. Therefore potential avalanche and snow-blowing areas were expressed as ratios of the area of the respective former glacier; these ratios are presented in Table 7.2. As all the avalanche ratios are relatively small, it seems that contributions from this source were relatively unimportant. With the exception of the Coire Dearg glacier, it is apparent from the snow-blowing ratios that all the former glaciers show the predominance of the potential contribution to accumulation by snow blown from the SW quadrant and that winds from between west and north would also provide a significant amount. The large SW and NW snow-blowing areas associated with the Stronchrubie glacier may well account for its low equilibrium firn line and probably for the existence of this glacier at all.

7.4 PALAEOCLIMATIC INFERENCES

It is clear from the figure and the tables presented in this chapter that, in the study area, Loch Lomond Advance glaciers with eastern and southern aspects were larger than those facing north and

GLACIER	AVAL. RATIO	SNOW-BLOWING RATIOS			
		SW	SE	NW	NE
QUINAG	0.214	0.106	0.023	0.094	0.085
COIRE DEARG	0.455	0.346	0.631	0.100	0.113
COIRE GORM	0.251	0.420	0.130	0.254	0.036
BEINN UIDHE	0.113	0.261	0.239	0.190	0.016
BEINN AN FHURAIN	0.074	0.210	0.119	0.077	0.050
STRONCHRUBIE	0.044	1.064	0.148	0.703	0.136
GLEN OYKEL	0.069	0.122	0.090	0.121	0.079

TABLE 7.2

Potential avalanche areas and snow-blowing areas expressed as ratios of glacier areas.

north-east, and possessed lower equilibrium firn lines. Similar situations were found on Skye (Sissons 1977b) and Rhum (Ballantyne & Wain-Hobson 1980), such a pattern being cited as evidence for heavier snowfall on the southern sides of the individual mountain groups, Sissons (1980a) inferring that the principal snow-bearing air streams came from the south. The evidence from Assynt therefore seems to support this view, with the larger glaciers on the southern and eastern side of the mountains. There is no reason to suspect that the recalculated equilibrium firn lines presented here will significantly change the trend of the regional firn line described by Sissons (1977a, 1980a) for northern Scotland as a whole.

The predominance of the potential contribution from snow blown onto glacier surfaces by winds from between south and west has been noted elsewhere (e.g. Sissons 1980a); the results of the present work strengthen this view. For many of the Loch Lomond Advance glaciers in the study area, winds from the NW quadrant might also have contributed significantly to glacial accumulation. In the Assynt area the inverse relationship between SW snow-blowing areas and firn line altitudes described by Sissons (1980a) and echoed by Cornish (1981) is only a weak one and would not exist at all if the values for the Stronchrubie glacier were discounted as exceptional. However, the sample size is too small to be representative of northern Scotland as a whole.

Multiple end moraines produced by the Stronchrubie and Beinn an Fhurain glaciers, and multiple lateral moraines formed by the Glen Oykel glacier, suggest that the margins of these glaciers fluctuated about their maximum positions during the Loch Lomond Stadial, as their mass balances tried to maintain steady states during conditions of differing accumulation and ablation. At present there is no evidence to suggest that widespread synchronous minor climatic ameliorations and deteriorations occurred that might have caused such a pattern of marginal glacial features during the stadial, enabling particular sets of features to be correlated from glacier to glacier. In Scotland, the lack of recessional moraines some distance up-glacier from the end-moraine complexes, with the exception of the 'ice-cored' moraine of the Beinn an Fhurain glacier (section 6.3 (e)) and another near Loch Skene in the central Southern Uplands (Sissons 1976, p. 106), suggests that rapid deglaciation occurred at the end of the stadial. A slightly different situation prevailed in the English Lake District, where during deglaciation "the great majority of the small glaciers retreated actively for 100-150 m and then decayed _in situ_, but some, at least, of the larger glaciers were active during much of the time their margins retreated" (Sissons 1980c, p. 13).

CHAPTER 8 : GLACIAL FRICTION CRACKS

8.1 INTRODUCTION

In addition to the linear abrasion features commonly seen on glaciated bedrock surfaces, other microerosional forms have been observed by past workers. They are often arcuate in plan, consisting of fine fractures and notches. They occur as individuals or in series. They have been collectively referred to as 'chattermarks' (Charlesworth 1957, vol. 1, pp. 248-249; Fairbridge 1968, p. 117), 'crescentic marks' (Flint 1957, pp. 62-64) and 'friction cracks' (Thwaites 1956, pp. 24-25; Embleton & King 1975, pp. 187-191; Embleton & Thornes 1979, p. 275). Few detailed studies of these features have been attempted: as these microerosional forms are abundant on the glaciated quartzite surfaces in the Assynt area, it was decided that they were worthy of a special study.

8.2 PAST WORK ON GLACIAL FRICTION CRACKS

One of the earliest papers dealing with these small-scale, glacially-eroded features was T.C. Chamberlain's monumental work on glacial striations (Chamberlain 1888). He recognised several types: 'chattermarks', 'jagged grooves', 'crescentic cross-fractures', 'jumping gouges' and 'lunoid furrows'. However, Chamberlain offered little more than a description of these features.

Gilbert (1906) concentrated on 'crescentic gouges', 'chattermarks' and 'crescentic cracks' (the latter being synonymous with Chamberlain's 'crescentic cross-fractures'), as these were seen to be the three principal types. A description of each was followed by a discussion of their possible mode of formation. Gilbert explained chattermarks as due to an essentially rhythmic motion of a boulder that was not firmly entrained in the ice, passing over brittle bedrock. The crescentic crack was seen as a result of tensile stress parallel to the rock-face. As the ice moved forward it tended to carry the bedrock along with it due to friction, perhaps increased by fine basal debris. Relative movement of any point of the bedrock surface "involves compression about the downstream side of the affected rock and tension about its upstream side, the magnitude of the stresses depending on the differential friction, and rupture ensuing when the tensile stress exceeds the strength of the rock" (Gilbert 1906, p. 304). Crescentic gouges were explained as a result of conoid fracture of the bedrock. Pressures normal to the bed (e.g. the weight of the block and the ice above the block), perhaps concentrated on one particular point on the rock surface by way of sand-sized particles trapped beneath the entrained block, and forces of differential friction between ice load and bed, were seen as the causative factors producing stress within the bedrock. When the strain limit of the bedrock was reached, a fracture was initiated, obliquely downwards. Tensile stresses above and compressive stresses below the wedge of rock above this primary fracture (ADC in Fig. 8.1(a)) caused a second fracture (FG in Fig. 8.1(a)) to form, completing the crescentic gouge when the now detached wedge of rock was removed. Although empirical in its approach, this paper remains to date one of the few detailed discussions of the possible ways in which these features were formed.

Lahee (1912) examined an exposure of quartzite that exhibited good examples of what he termed 'crescentic fractures' (synonymous with Gilbert's 'crescentic cracks'). He agreed with Gilbert's suggestions

Fig. 8.1 Past views on the formation of certain glacial friction crack forms.

(a) G.K. Gilbert's explanation of the formation of crescentic gouges (after Gilbert (1906)).

Key: A-E Original rock surface.

A-B Rock surface deformed by entrained block.

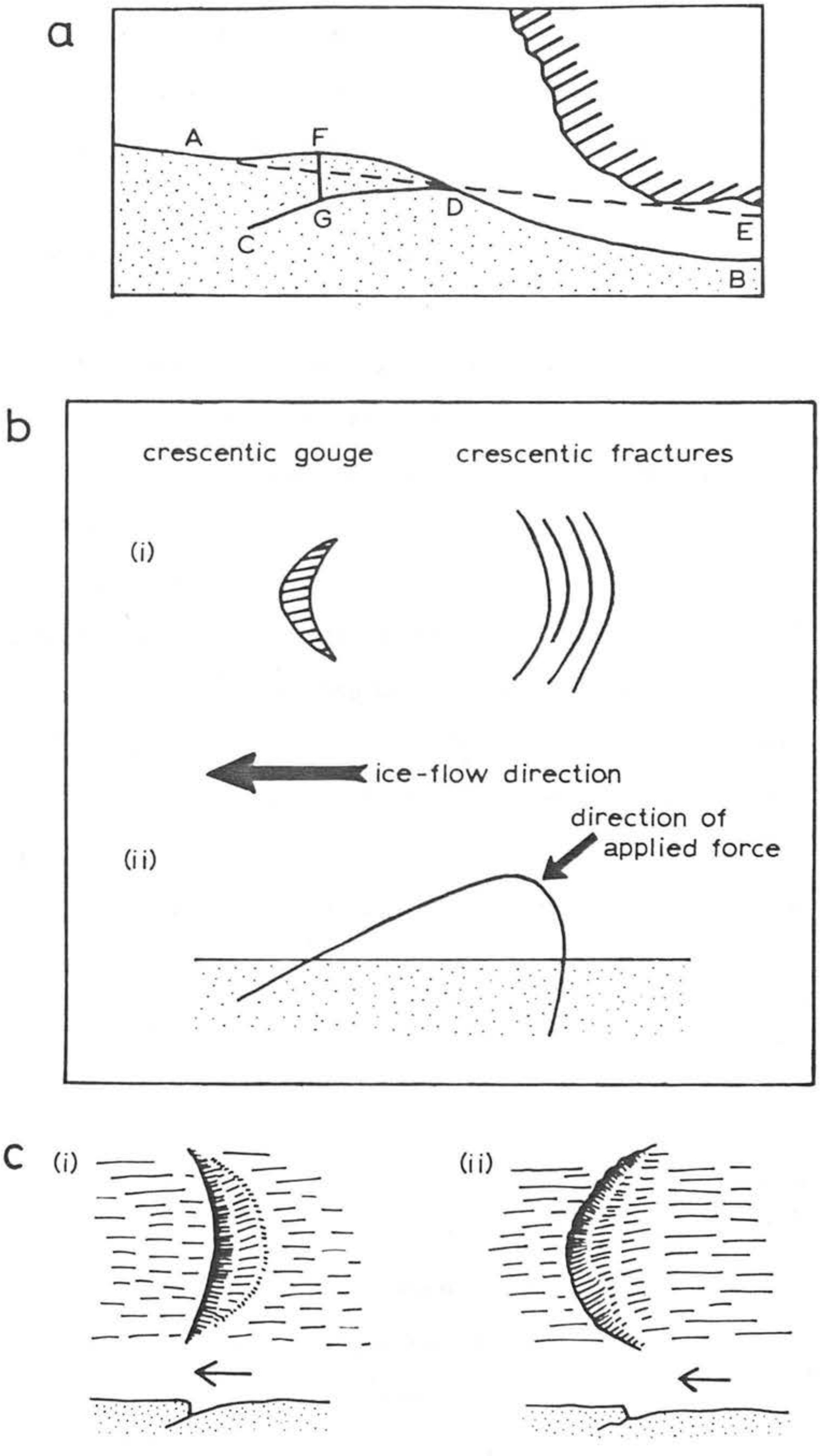
D-C Primary fracture.

F-G Secondary fracture.

(b) E. Ljunger's views on the relationship of crescentic gouges to crescentic fractures. (i) Plan view; (ii) cross-sectional view, showing the 'conoid of percussion (adapted from Dreimanis (1953)).

(c) Diagrams of a crescentic gouge and a lunate fracture (after Harris (1943)).

(diagram overleaf)


a
A
F
D
G
C
E
B
b
crescentic gouge
crescentic fractures
(i)
ice-flow direction
direction of applied force
(ii)
c
(i)
(ii)

as to the creation of these forms, except that he argued for the operation of a pointed or edged tool rather than "pockets of sand" entrained in the ice. His reasons for this were fourfold:

(i) the increase in the number of fissures towards the axes of series of these fractures;

(ii) the petering out of these fractures both downward and laterally, away from the axes;

(iii) the frequent association of striae and grooves with many of the crescentic fracture series;

(iv) the fact that the fractures are hyperbolic rather than truly crescentic.

Lahee also suggested that there is a preferential positioning of these fractures at or near the crests of the proximal side of bedrock protuberances. The even spacing between the fractures in a series (also reported by Gilbert 1906, pp. 312-313) was seen as representing a rhythmic build-up and release of tensile stresses, with a sort of 'stick-slip' motion of the entrained block.

Ljunger (1930) applied his own classification to these features, adding a new form to those previously recognised - the conchoidal fracture ('Muschelbruch'), with a concave-upward fracture plane. He adopted Gilbert's ideas on conoid fracture to explain the formation of crescentic gouges and crescentic fractures, which he saw as being either side of an obliquely-positioned conoid of percussion (Fig. 8.1(b)). However, two points cast doubt on Ljunger's views.

(i) As Gilbert had commented previously, it is difficult to see how a block entrained in glacier ice can effect a blow on the bedrock to create such a conoid of percussion.

(ii) Crescentic gouges and crescentic fractures are rarely seen in close association in the way depicted in Fig. 8.1(b).

In view of these points, it is difficult to accept Ljunger's ideas on the formation of such features.

Harris (1943) introduced the collective term 'friction crack' to describe those of the above forms that exhibit a primary fracture. His classification therefore excluded chattermarks, but included a new form - the 'lunate fracture' - which is similar in morphology to the crescentic gouge except that the 'horns' of the crescent point the other way (Fig. 8.1(c)). Because of this, Harris drew attention to the fact that the direction in which the horns point is an unreliable indicator of the direction of former ice flow. In this respect, consideration of the direction of dip of the primary fracture of all types of friction crack was seen as more useful, the fracture dipping down-ice : this was apparently confirmed by the work of Okko (1950) in Finland. Harris attempted to reproduce the various crack forms using different tools on a selection of materials. Forms analogous to crescentic fractures were easily produced on a variety of surfaces, but crescentic gouges were produced only on glass, using a sharper tool and more pressure than was needed to produce crescentic fractures. However, on his own admission, Harris did not attempt to reproduce conditions to scale in respect of time, pressure, hardness or character of material or cutting tool, in any of his experiments.

MacClintock (1953), citing the work of Preston (1921), suggested that the different friction crack forms may be related, in part at least, to whether or not the tool producing the features was held fast in the ice. Preston's work on the fracturing of optical glass using ball-bearings had produced forms analogous to crescentic fractures when a clamped ball-bearing had been forced across the glass surface. However, when the ball-bearing was not held so fast, and allowed to rotate whilst a force was still exerted from above, forms similar to the primary fractures of crescentic gouges were produced, although a secondary fracture was not present.

The results of a study of friction cracks in Ontario caused Dreimanis (1953) to question Harris's generalisation that their primary fractures dipped with the forward direction of former ice movement. The majority of crescentic fractures in the study area were shown to dip against the former ice-flow direction that was indicated by the distribution of glacial erratics. Following on from this, Dreimanis attempted to see if there was any relationship between the observed dip of the fractures and the structure of the rocks on which they were found. He also tried to determine if there were any other influences on the distribution of friction cracks, such as the slope of the rock surface with respect to ice movement. He concluded that rock structure (in this case schistosity) does influence the dip of the fracture, increasing the dip in the examples studied (although his method of obtaining such measurements was clearly unsatisfactory, especially in view of the form of the fracture as indicated by later work (Johnson 1975)). As Lahee had noted, friction cracks were found in greatest abundance near the crests of rock protuberances, on their proximal side.

Thwaites (1956) hypothetically related the formation of crescentic fractures and crescentic gouges to the amount of the resultant force, from the motion of the glacier and the effective normal pressure, acting at a point on brittle bedrock. However, he presented little detailed evidence to support his ideas.

Andersen & Sollid (1971) discovered in southern Norway what they considered to be reversed crescentic gouges, i.e. "forms with the morphological characteristics of crescentic gouges, turned 200° (sic) in relation to normal gouges" (ibid., p. 18) and thus not to be confused with lunate fractures. An explanatory diagram (ibid., p. 17) only serves to confuse the issue as the legend appears incorrect, and a photograph (ibid., Fig. 20(b)) purporting to show reversed crescentic gouges illustrates features resembling a glacially 'plucked' rock surface. However, another photograph (ibid., Fig. 20(a)) does indeed show crescentic gouges that appear to be reversed, on the distal side of an ice-moulded rock protuberance.

An unpublished Ph.D. thesis by C.B. Johnson (1975) is one of the most detailed studies of friction cracks to date. He characterised the different types of observed friction cracks and investigated the suggested mechanics of formation of each one. Of his thesis, he said (p. 2) "a specific goal is to determine in what manner the kinematics of the entrained block is reflected in the geometry of the crack" in order to test the relationship between the cracks and the direction of ice movement. It must be pointed out that Johnson described these features collectively as 'glacial arcuate abrasion cracks'. Such a term is unsuitable since the word abrasion implies the scraping-off of material from one surface by another, which in itself is not the cause of the features in question. 'Friction cracks' will therefore be retained as the collective term.

Johnson recognised four types of friction crack: 'Type 1' cracks, analogous to crescentic fractures; 'Type 2' cracks, analogous to crescentic gouges; 'Type 3' cracks, a previously unrecognised group similar to the primary fracture alone of crescentic gouges (no secondary fracture being present); and 'Type 4' cracks, which are completely circular cracks mentioned by Gilbert (1906) and referred to as 'ring cracks'. Similar experiments to those carried out by Preston (1921), using steel ball-bearings and ordinary window-glass, gave similar results: the different forms of crack depended on the kinematic state of the ball-bearing 'tool'. Johnson then related each of the four main friction crack groups to one of the three experimentally-produced crack types (i.e. from the 'tool' allowed to roll with downward pressure, held rigid and pushed, and pushed straight down into the substrate) on the basis of morphological similarities. Despite careful experimentation and highly quantified work, two questions were left unanswered by Johnson's work.

(i) What is the origin of the secondary fracture of crescentic gouges, which Johnson was unable to produce successfully in his experiments?

(ii) Is the action of steel on glass an adequate analogy for ice-entrained rock particles and bedrock?

Nevertheless, Johnson felt able to suggest the mechanisms by which these arcuate features might be formed in a subglacial situation (discussed in section 8.4.2, below). Although he looked briefly at the influence of factors other than the kinematic state of boulders in the ice (e.g. the presence of meltwater at the base of the ice, and the effect of the weight of the entrained block), he made few inferences from the results of his work. He suggested two areas where future research should be carried out:

(i) the analysis of ice - block interactions at the base of a glacier;

(ii) quantitative analysis of the spatial distribution of the basic friction crack types, in order to determine the exact positioning relationship of features of varying size and form with respect to large and small topographic features.

He went on to say:

> "Ultimately, if the mode of block motion associated with each ... (friction crack) ... type, the details of ice - block interactions and the resultant relationships between block motion and far field velocity patterns in the surrounding ice are all known, then it should be possible to evaluate the field observations of the spatial distribution patterns of the cracks with respect to providing additional details about the movement of ice in the basal layer of a glacier as it moves over an undulating surface."
>
> (Johnson 1975, p. 229)

A recent article by Thorp (1981) has briefly discussed friction cracks on glaciated surfaces in part of the western Grampian Highlands of Scotland. Thorp measured the widths (i.e. from 'horn' to 'horn') of a number of friction cracks and found that crescentic fractures ranged from 2-40 cm (although the majority were less than 10 cm), and crescentic gouges varied from 3-50 cm (with the majority falling between 10 cm and 20 cm). He noted that at any one site the orientations of crescentic gouges varied: he attributed this to local geological structure, although no quantitative evidence was presented to support this view. Thorp recognised reversed forms of both crescentic gouges and crescentic fractures, although examples of the latter were only seen at six localities amongst the numerous sites he studied. Thorp stated that in the localities where such reversed friction crack forms were found it was inconceivable that two directions of former ice flow could be represented. He argued that an earlier phase of markings would be unlikely to survive a later ice advance. However, this overlooks the fact that crescentic fractures commonly extend several centimetres into the bedrock and that removal of a layer of rock by abrasion would expose cracks similar in plan to the original features. In view of their scarcity, Thorp's interpretation of crescentic fractures that are apparently reversed in form should be viewed with some caution.

8.3 SOME PRELIMINARY FIELD OBSERVATIONS

The writer made the following observations about friction cracks in the Assynt area during an early stage of fieldwork. The terminology used in this section follows that of Embleton & King

(1975, Fig. 6.3).

(i) Unequivocal friction cracks were only found on quartzites. This is partly due to the tendency of this hard, crystalline rock type towards brittle fracture, and partly due to the resistance of quartzite to chemical and small-scale mechanical weathering (i.e. granular disintegration).

(ii) Examples of all types of friction cracks, i.e. corresponding to Johnson's types 1, 2, 3 and 4, were recognised. Crescentic gouges and crescentic fractures comprised the majority of the friction cracks studied. Ring cracks were very rare and were all poor examples, the circular crack pattern never being complete. Crescentic fractures often occurred in series (i.e. three or more aligned cracks); crescentic gouges were sometimes present in series, but more often they occurred individually. Crescentic gouges were found on glaciated surfaces with a variety of aspects in relation to former ice flow, but crescentic fractures tended to be more numerous near the crests of bedrock prominences, on their proximal sides.

(iii) A wide variety of shapes of crescentic gouges was noted, ranging from classic 'crescentic gouge' forms through to 'lunate fractures'. For this reason it is suggested that lunate fractures should be considered merely as a sub-group of crescentic gouges. Composite forms were also noted where two or more crescentic gouges were joined to form one feature.

(iv) When their basic morphology is considered, crescentic gouges and crescentic fractures seem to conform in orientation at least approximately with the former direction of ice movement suggested by the distribution of glacial erratics. This apparent relationship needs to be tested by systematic measurement. However, reversed crescentic gouge forms do exist, albeit rarely. Some attempt must therefore be made to ascertain whether such features favour particular positions on glaciated surfaces, and to what extent they were produced contemporarily with normal forms. No instances of reversed crescentic fractures were found.

(v) The structure and texture of the bedrock seem to influence the detailed orientation and degree of development of certain of the features. Quantification of these apparent relationships is therefore necessary if the value of friction cracks as directional indicators of former ice flow is to be correctly assessed.

(vi) A sample of rock collected from the quartzite dipslope of Canisp fortuitously showed a series of crescentic fractures in cross-section. The fractures clearly dipped backwards, against the direction of former ice movement. This supports the argument (Dreimanis 1953; Johnson 1975) that Harris's views on the dip of the primary fractures of such features are invalid.

(vii) Along the top edge of the distal side and at the very crest of the ice-moulded rock surfaces, lunate fractures and crescentic fractures appear to be most numerous. It is possible that such features play a large part in the glacial 'plucking' process, weakening bedrock and even loosening chips which are subsequently removed by the ice. However, it is difficult to see whether they are a cause or an effect of this erosion process.

8.4 SYNTHESIS OF FACTS, AND INFERENCES

Previous work has been largely empirical and unquantitative. Various classifications of friction cracks have resulted in a confused nomenclature. Certain aspects of friction crack morphology seem to be useful if one wishes to determine the direction of former ice flow in a glaciated area where such cracks exist. On the other hand, there have been few attempts to elucidate the manner of their formation and the nature of the forces involved, and the extent to which their distribution over a bedrock surface might be explained by topographic and structural influences, as well as glaciological controls, has received little attention.

A synthesis of the known facts about friction cracks can be sub-divided into

(i) a presentation of a classification of such features and a discussion of the suitability of nomenclature;

(ii) a discussion of the possible modes of formation of the different crack types, and inferences that can be based on this work;

(iii) a consideration of the distribution of friction cracks and possible factors influencing that distribution.

8.4.1 A classification of friction cracks

It is appreciated that, where a confused nomenclature already exists, further redefinition of terms seldom helps. Nevertheless, the writer is of the opinion that several terms applied to friction cracks are imprecise and many past classifications unnecessarily complex.

Johnson's classification is here used as a basis since it is the simplest proposed to date and the most appropriate in view of his work into crack morphology and mechanics of formation. However, it is thought appropriate to change those of his descriptive terms that are considered inaccurate.

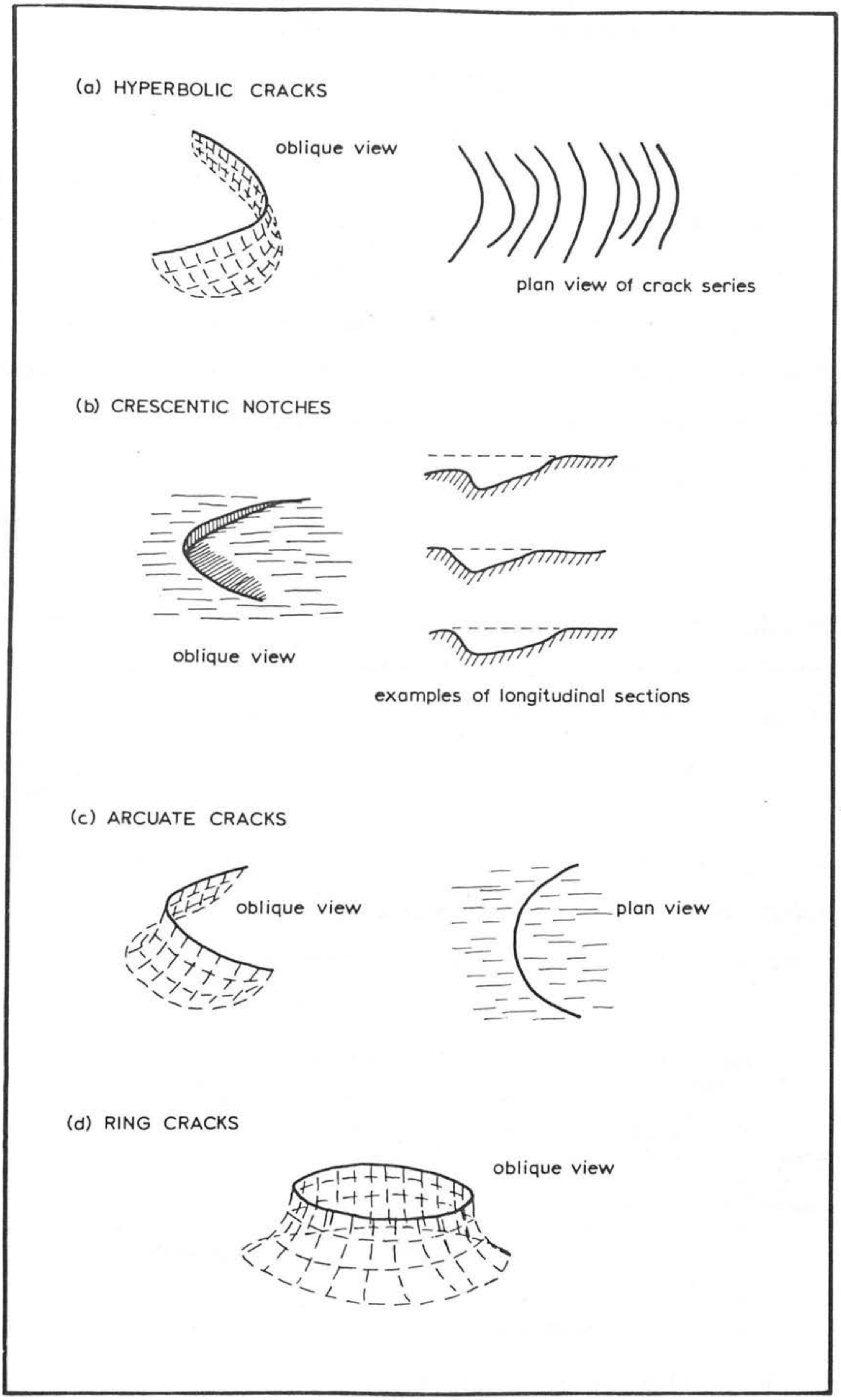
Four types of friction crack have been recognised:

(a) Type 1. In plan view consisting of fractures, with straight limbs either side of a central curvilinear segment, concave forward in the direction of former ice movement (Fig. 8.2(a)). In vertical section, they possess a steeply dipping segment, often (but not invariably) inclined towards the forward direction, replaced beneath by a curvilinear or straight segment dipping and flaring out rearwards. This type of friction crack is most commonly found as one of an aligned series. They have been described by past workers as 'crescentic fractures' or 'crescentic cracks'. However, they are not strictly crescentic (= 'moon-shaped') and it is therefore suggested that 'hyperbolic crack' is a more appropriate descriptive term.

(b) Type 2. These features consist of a primary fracture of forward inclination and a secondary curviplanar crack most often of forward convexity. The intersection of the two cracks has resulted in the release of a lunate wedge of rock (Fig. 8.2(b)). Isolated individuals of this type of friction crack are common, but they do occur sometimes in series. In some cases it is possible to divide the primary crack into two distinct regions. The secondary crack intersects the primary fracture at a high angle (Johnson (1975) observed a range of 70°-130°) and clearly forms after it. These features have usually been referred to as 'crescentic gouges', a term that Gilbert considered was inappropriate as "the word 'gouge' connotes a process of formation analogous to a

Fig. 8.2 A classification of glacial friction cracks (adapted from C.B. Johnson (1975)).

(diagram overleaf)


(a) HYPERBOLIC CRACKS
oblique view
plan view of crack series
(b) CRESCENTIC NOTCHES
oblique view
examples of longitudinal sections
(c) ARCUATE CRACKS
oblique view
plan view
(d) RING CRACKS
oblique view

chisel" (Gilbert 1906, p. 313). It is therefore suggested that these forms are better termed 'crescentic notches'.

(c) Type 3. Un-named by Johnson, it is suggested that these friction cracks should be called 'arcuate cracks'. In plan view they consist of a single fracture, approximating an arc of a circle, convex forward (Fig. 8.2(c)). They do not possess the straight limb segments of the hyperbolic cracks (type 1, above). There is limited information on the form of these arcuate cracks in vertical section, but they appear to intersect the rock surface at right angles, then flare outwards in a forward direction. This crack type is relatively rare.

(d) Type 4. Recognised initially by Gilbert (1906), these 'ring cracks' have a circular or elliptical surface trace. The crack intersects the bedrock surface at right angles and then rapidly flares outward in all directions (Fig. 8.2(d)).

8.4.2 Mode of formation

Johnson (1975) saw friction cracks as caused by the direct contact of a glacially-entrained block with the bedrock floor of the glacier. He stated (p. 48):

> "During contact, the block exerts on the bed a force, the magnitude and orientation of which depend upon both the block's weight and the surface forces acting on it as a result of ice - rock block interactions. As a consequence of the small size of contact, the bed experiences a concentrated loading which generates an intense, localised deformation of the rock in the vicinity of the contact. In certain instances the deformation state is sufficiently intense to activate a brittle fracture event, which results in the formation of a macroscopic crack."

This model of actual block-bedrock contact was based on the marked geometrical similarity between the observed friction cracks and cracks produced experimentally.

The results of experiments using steel ball-bearings and glass, mentioned above, suggested that the different friction crack morphologies reflect, at least in part, the differing kinematic states of the contacting rock blocks. The results of Johnson's experiments are presented schematically in Fig. 8.3.

If the suggestions as to the mode of formation of the different friction cracks as presented in Fig. 8.3 are acceptable, bearing in mind that steel on glass may prove not to be analogous to the subglacial situation, the following points can be made regarding their glaciological significance.

(i) Analysis of the orientation of friction cracks can give the local direction of former ice flow. In all cases the movement vector is transverse to an individual crack and parallel to the axis of a crack series. One or more characteristics of each crack type is diagnostic of the direction of former ice movement. Obviously, the orientation of friction cracks is strongly influenced by the flowage of ice over the local bedrock topography, so it is necessary to measure a large number of friction crack orientations to ascertain regional ice-flow directions correctly.

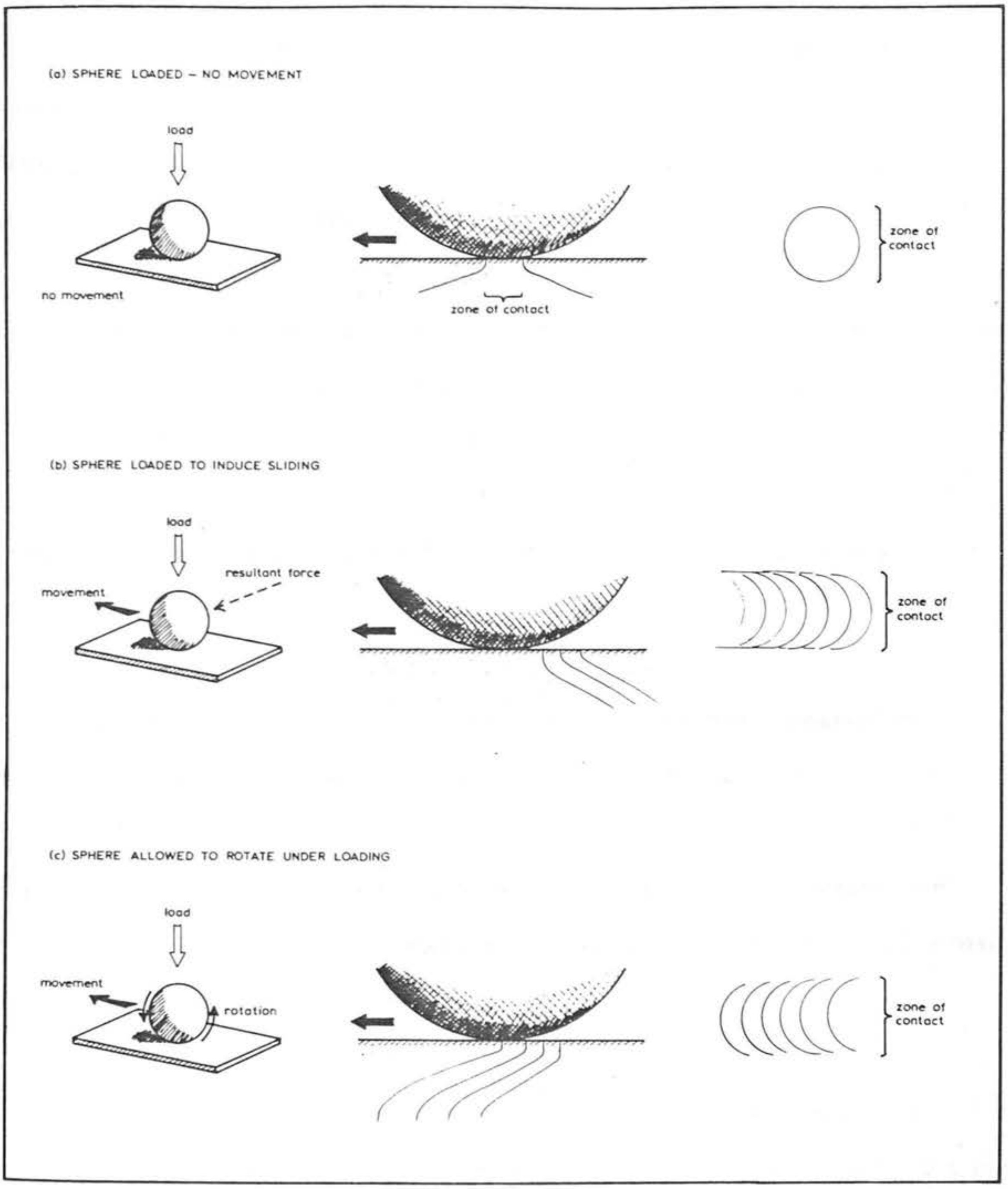
(ii) Glacially-entrained blocks of suitable size and shape, and sufficiently thick glacier ice, must have been present to provide the forces required to produce friction cracks. However, there are at present no reliable estimations as to the magnitude of these parameters.

(iii) Subglacial transportation of some blocks involving rolling or rotation is apparent as well as the generally accepted dragging

Fig. 8.3 Summary of C.B. Johnson's experimental production of cracks on glass, using a steel ball-bearing (adapted from Johnson (1975)).

(N.b. left-hand diagrams are oblique views, central diagrams are cross-sectional views, and right-hand diagrams are plan views of the resultant crack forms.)

(diagram overleaf)



(a) SPHERE LOADED – NO MOVEMENT
load
no movement
zone of contact
zone of contact
(b) SPHERE LOADED TO INDUCE SLIDING
load
resultant force
movement
zone of contact
(c) SPHERE ALLOWED TO ROTATE UNDER LOADING
load
movement
rotation
zone of contact

of blocks across the glacier bed.

8.4.3 Distribution of friction cracks

The distribution of friction cracks noted by past workers appears to be spatially non-random. The factors influencing the distribution are probably geological, glaciological and topographical in character. They include:

(i) the distribution of rock types that are suitable for the formation and long-term preservation of friction cracks, i.e. hard and brittle rocks that are not prone to weathering;

(ii) the availability of rock fragments of suitable size and shape to act as tools;

(iii) the relationship of both micro- and macro-scale topographic features to stress fields at the base of the glacier;

(iv) the complex interrelationships of thermal regime, presence or absence of basal meltwater, ice thickness, ice velocity and other glaciological parameters.

The observations of Lahee (1912), Dreimanis (1953) and Johnson (1975) suggest that hyperbolic cracks tend to be concentrated near the crests of bedrock protuberances, on their proximal side, whereas crescentic notches are more widespread in their pattern of distribution. There are, however, no quantitative data on this topic.

8.5 ANALYSIS OF THE ASSYNT FRICTION CRACKS

8.5.1 Objectives and methods

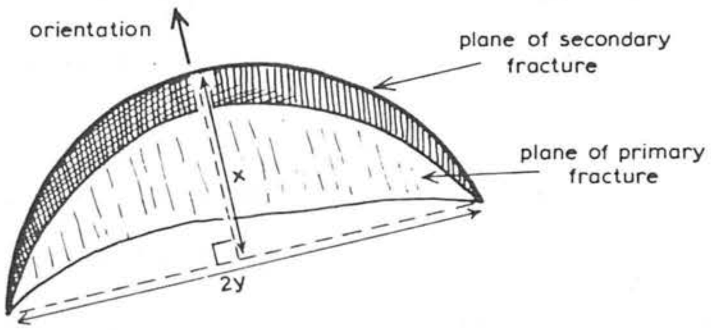
A major aim of the present study was to improve upon the sparse amount of quantified information relating to glacial friction cracks. The author was unsure as to what particular properties of the various friction crack types would prove most useful in the subsequent study, so the fieldwork was designed to collect as much information as possible. Early attempts to measure specific parameters *in situ* were hampered by adverse weather conditions and proved to be slow and laborious. It was decided that a better method of analysis would be to trace a large number of friction cracks from the chosen sampling sites directly onto pre-orientated transparent, weatherproof, melamine drawing film. At a later date measurements could be taken from the tracings, which could be orientated if necessary.

The main objectives of the study were to attempt to characterise the predominant types of friction crack and to determine whether or not they are reliable indicators of former ice-flow direction. Since crescentic notches and hyperbolic cracks were the types best represented in the Assynt area, measurements were restricted to these. Three measurements were taken from each crescentic notch: the 'horn'-to-'horn' measurement, the maximal distance from this line to the edge of the secondary crack (*i.e.* 2y and x respectively (Fig. 8.4(a)) and the orientation of the feature measured perpendicular to a line joining the two 'horns'. Five measurements relating to the size and shape of the hyperbolic cracks were taken (*i.e.* α, AB, AC, BC and z in Fig. 8.4(b)) as well as two orientation measurements: the first along an imaginary line bisecting angle α, and the second perpendicular to line BC (Fig. 8.4(b)). These two orientation measurements were taken because

Fig. 8.4 The geometry of a crescentic notch and a hyperbolic crack.

(diagram overleaf)

(a) The geometry of a crescentic notch



(b) The geometry of a hyperbolic crack

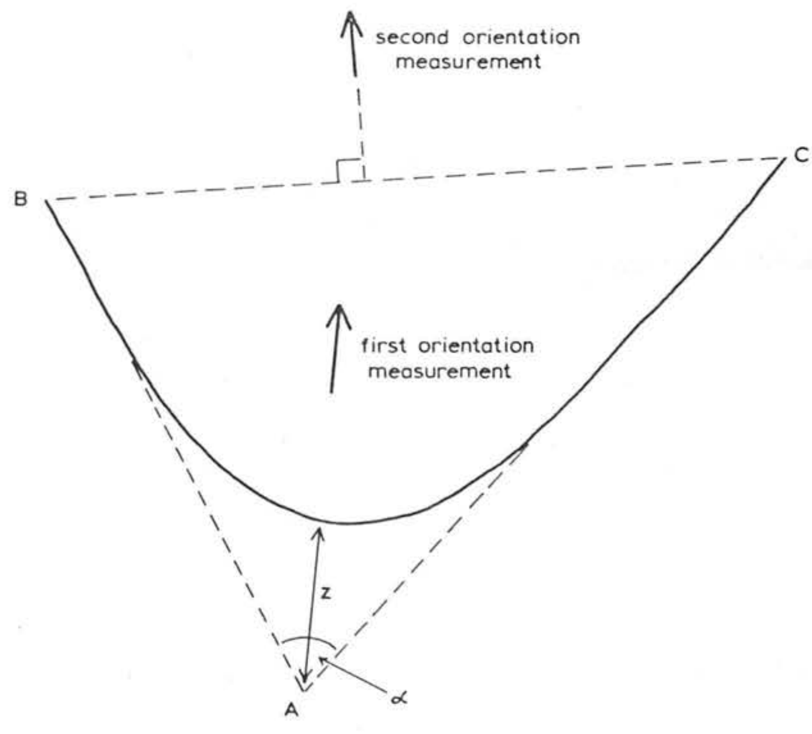


Fig. 8.5 Location of the sampling sites for the study of glacial friction cracks in the field area.

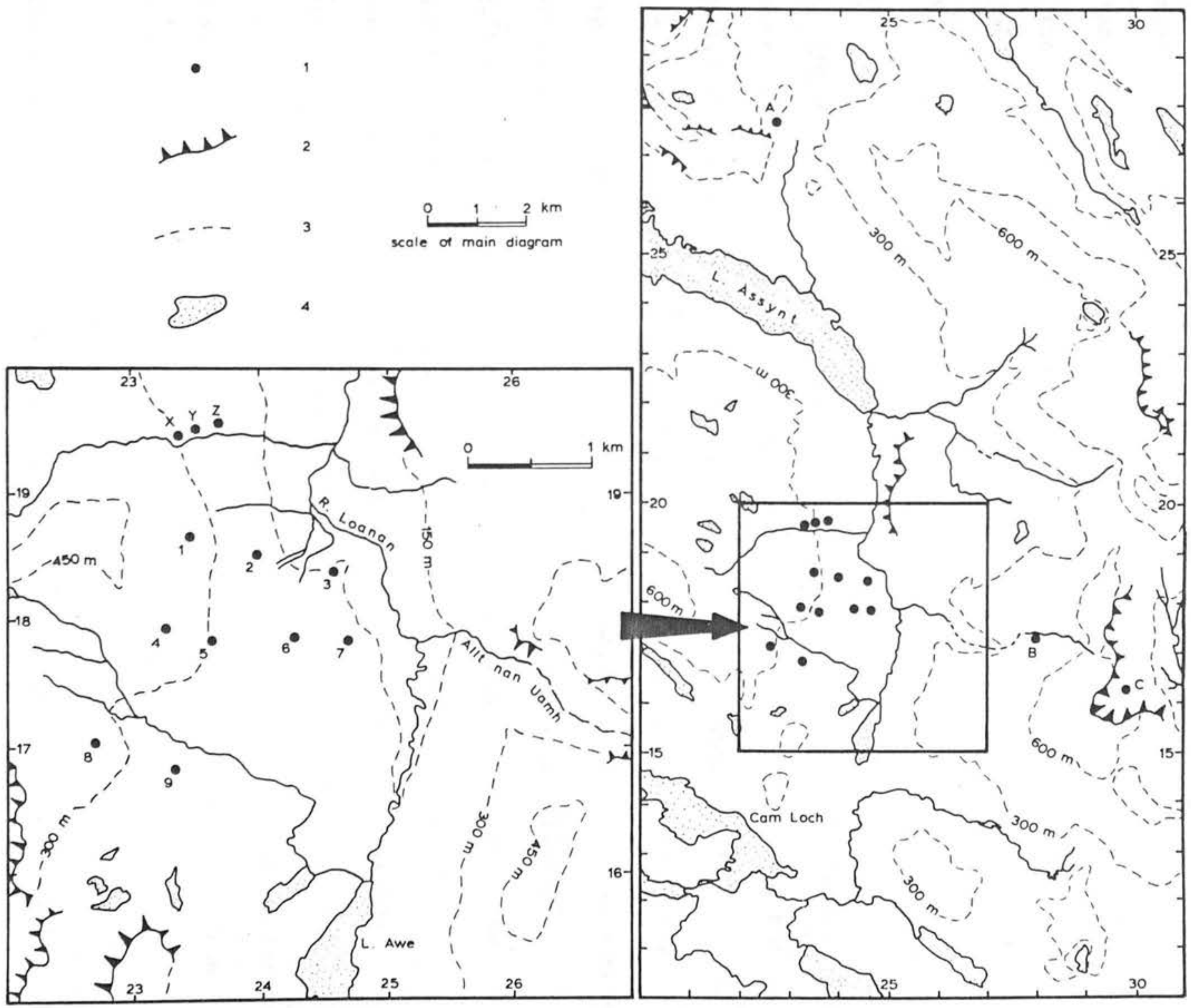
Key: 1. Sampling site.

2. Free faces.

3. Contours at 300 m intervals on the main diagram, and at 150 m intervals on the inset.

4. Main lochs.

(diagram overleaf)



1
2
3
4
0 1 2 km
scale of main diagram
X Y Z
1 2 3 4 5 6 7 8 9
0 1 km
R. Loanan
150 m
450 m
300 m
Allt nan Uamh
L. Awe
A
B
C
L. Assynt
Cam Loch
600 m

it was apparent from field observations that many hyperbolic cracks had 'horns' of different lengths, giving the crack an asymmetrical appearance which might reflect either ice movement or the direction of maximum applied force. During the study 1,370 friction cracks were analysed (977 crescentic notches and 393 hyperbolic cracks) from twelve sites on the eastern slopes of Canisp, and three others from elsewhere in the field area (Fig. 8.5).

8.5.2 Results

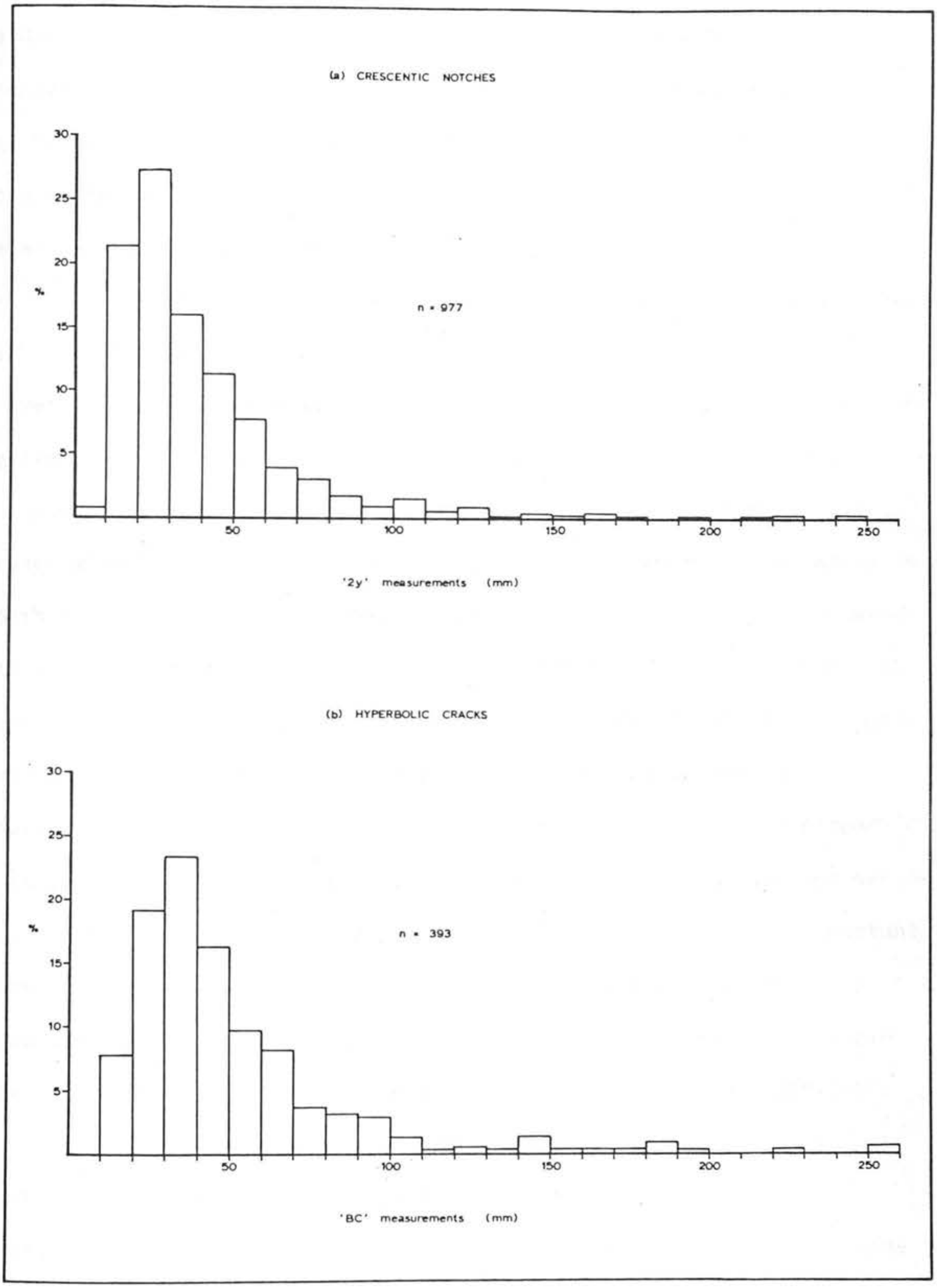
(a) Characterisation of the friction cracks. The crescentic notches ranged in size (i.e. from 'horn' to 'horn') from 7 mm to 1,010 mm, but the latter measurement was exceptionally large. A histogram of the size of the 2y-measurements for the total sample of crescentic notches (Fig. 8.6(a)) shows a unimodal distribution: 64.4% of the 2y-measurements fall between 10 mm and 39 mm, 27.2% occurring in the modal class of 20-29 mm. The hyperbolic cracks ranged in size (again from 'horn' to 'horn') from 12 mm to 296 mm. A histogram of the size of the BC-measurements for the total sample of hyperbolic cracks (Fig. 8.6(b)) shows that, on average, they are a little larger than crescentic notches: 58.8% of the BC-measurements fall between 20 mm and 49 mm, 23.4% occurring in the 30-39 mm modal class. No meaningful patterns were apparent from an analysis of friction crack size from site to site.

As the secondary fracture of crescentic notches in plan view fairly closely approximates the arc of a circle, it was decided that the radius of curvature of the nearest-fit circle would adequately characterise their shape. This radius of curvature was calculated for each crescentic notch from the equation $r = \dfrac{x^2 + y^2}{2x}$

Fig. 8.6 (a) Histogram of the 2y-measurements of crescentic notches from the Assynt area. (N.b. 21 measurements (2.2%), ranging from 264 mm to 1010 mm, are not included in the histogram due to problems of scale.)

(b) Histogram of the BC-measurements of hyperbolic cracks from the Assynt area.

(diagram overleaf)


(a) CRESCENTIC NOTCHES
30
25
20
%
15
10
5
n = 977
50
100
150
200
250
'2y' measurements (mm)
(b) HYPERBOLIC CRACKS
30
25
20
%
15
10
5
n = 393
50
100
150
200
250
'BC' measurements (mm)

the derivation of which is given in Fig. 8.7(a). The range of r-measurements was from 3.5 mm to 1,082.6 mm. The histogram in Fig. 8.7(b) shows that 49.6% of the total sample of crescentic notches had a radius of curvature of between 11 mm and 30 mm, 30.1% occurring in the modal class of 11-20 mm.

Hyperbolic cracks possess a more complicated geometric shape, which is difficult to characterise adequately. With many different parameters having been measured, linear regression of one against another highlighted only relationships that are largely as one might expect: z is positively correlated with BC and negatively correlated with α. Perhaps slightly more surprising is the fact that, where a site exhibits hyperbolic cracks of a wide range of sizes, BC is negatively correlated with α, implying that hyperbolic cracks which are large from 'horn' to 'horn' tend to have large 'limbs' separated by a relatively small angle. However, at sites where large hyperbolic cracks were absent such a relationship is only very weak or non-existent. Site 2 on the slopes of Canisp possesses hyperbolic cracks of a wide range of sizes and serves to illustrate these relationships (Fig. 8.8). One of the most important elements of the shape of the hyperbolic cracks is the angle α. Fig. 8.9 shows that 38.7% of the total sample of hyperbolic cracks had an angle between 91° and 110°, 21.3% occurring in the modal class of 101°-110°.

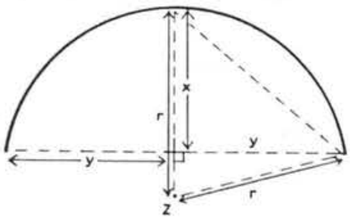
A further point of interest is the existence of friction cracks formed beneath Loch Lomond Advance glacier ice in the area (Figs. 8.10 and 8.11, sites X, Y and Z; also site C, not shown in these figures). This indicates that friction cracks can be formed beneath relatively thin glaciers; maximal ice thicknesses at sites X, Y and Z, calculated from the reconstructed surface of the Stronchrubie glacier

Fig. 8.7 (a) Derivation of the equation for determining the radius (r) of the best-fit circle for the secondary fracture of crescentic notches.

(b) Histogram of r-measurements of crescentic notches from the Assynt area. (N.b. the bar on the far right of the histogram represents 29 measurements, ranging from 161 mm to 1082 mm.)

(diagram overleaf)

(a)



$$r^2 = (r - x)^2 + y^2 \quad \text{(by Pythagoras)}$$

$$\therefore\ r^2 = r^2 + x^2 - 2rx + y^2$$

$$\therefore\ 2rx = x^2 + y^2$$

$$\therefore\ r = \frac{x^2 + y^2}{2x}$$

Z = unknown centre of circle whose arc closely approximates curvature of secondary crack of crescentic notch

r = radius of best-fit circle

x, y = measurements taken from crescentic notches

(arc) = secondary crack of crescentic notch

(b)

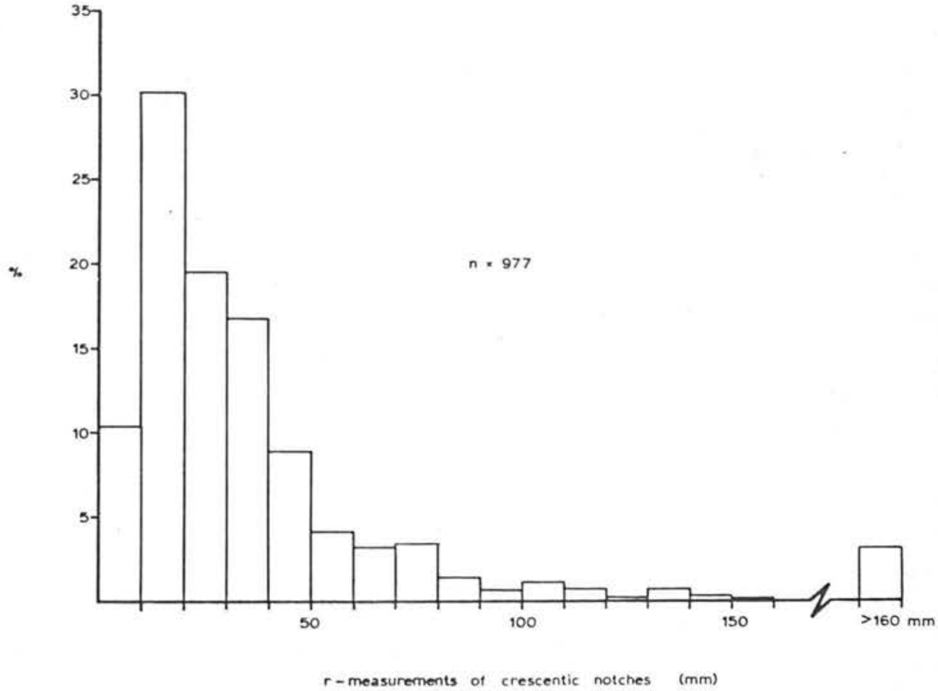
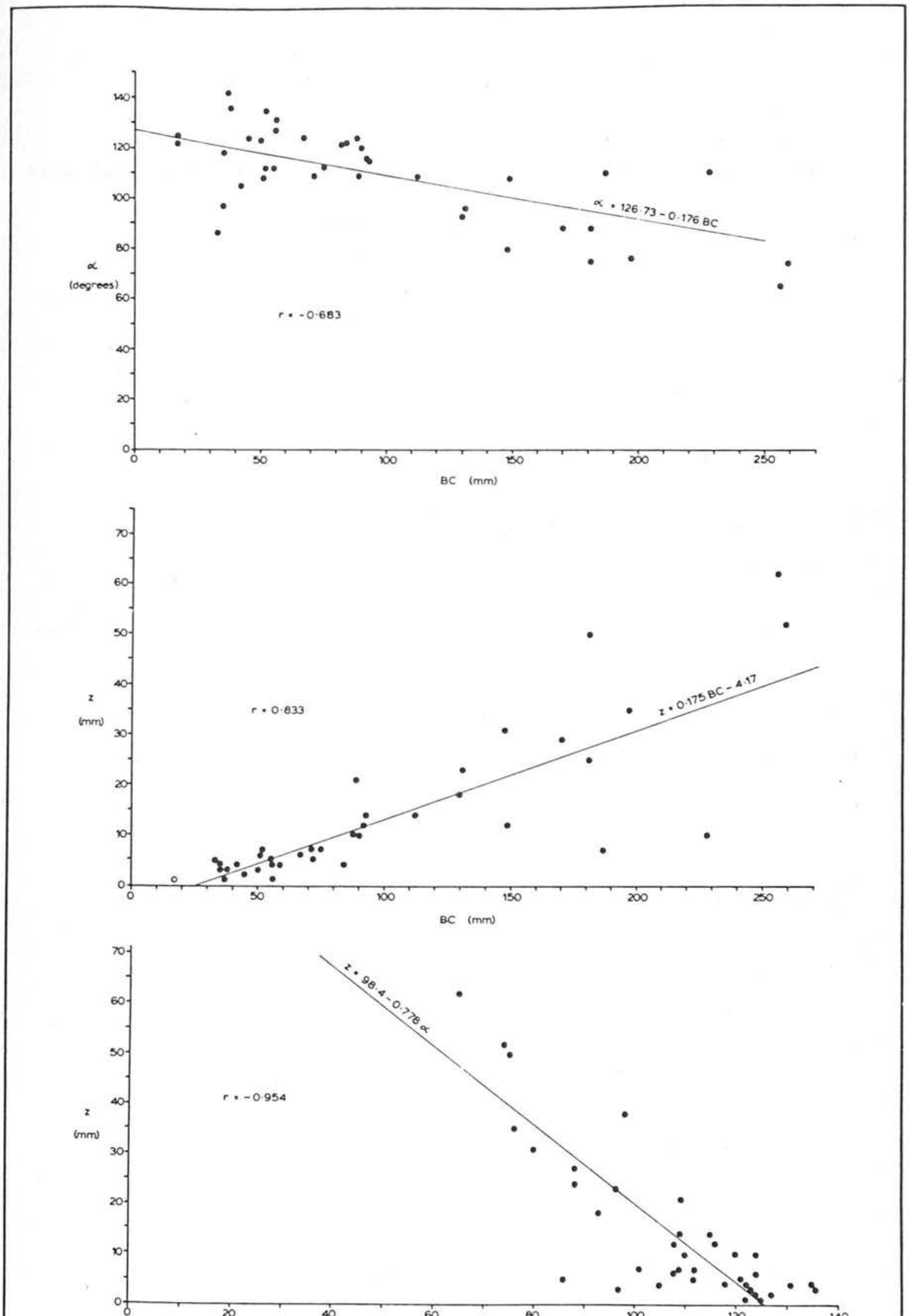


Fig. 8.8 Graphs showing the relationship between certain of the hyperbolic crack parameters at Site 2 on the quartzite dip-slope of Canisp.

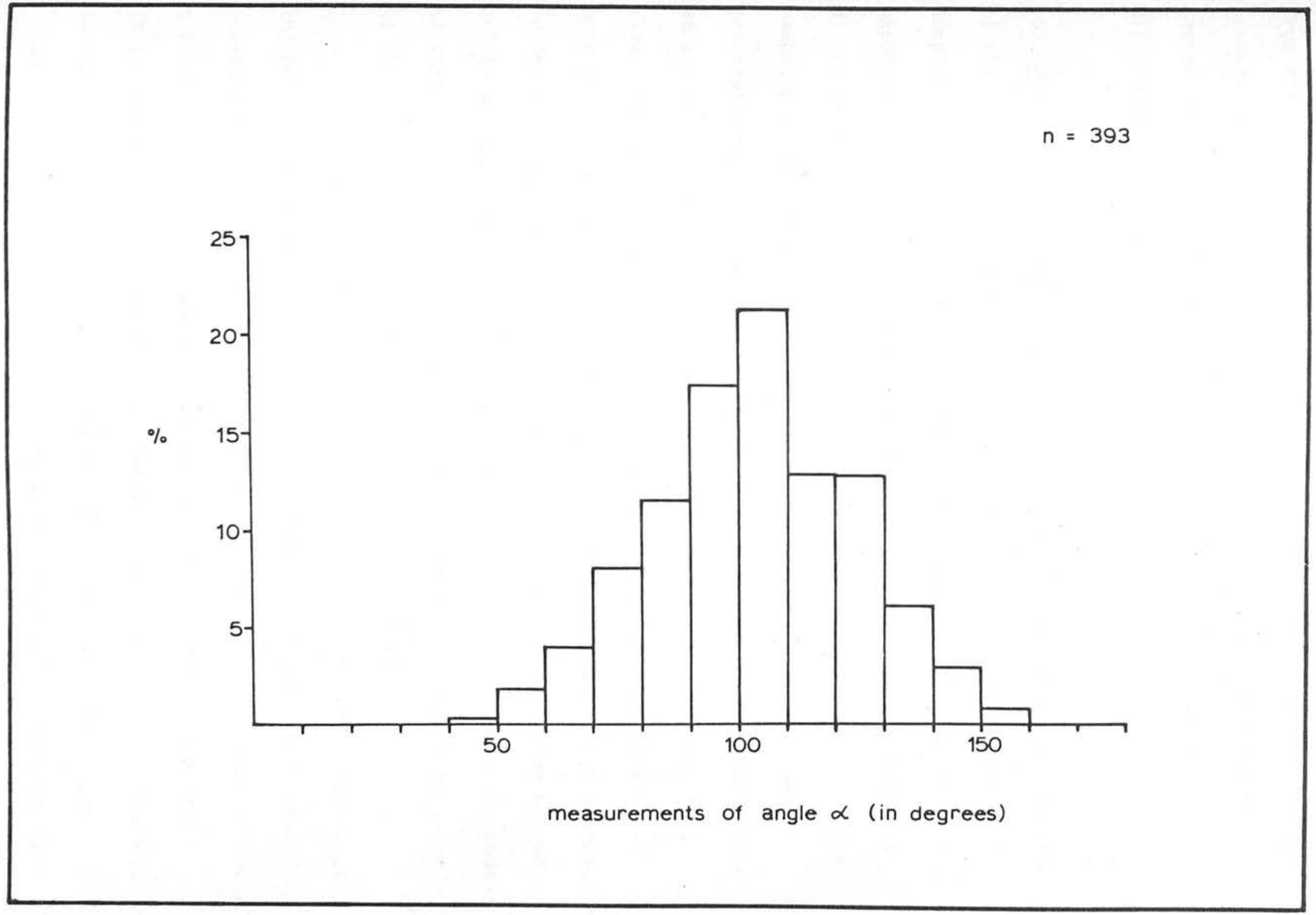
(diagram overleaf)



α = 126·73 − 0·176 BC
r = −0·683
α (degrees)
BC (mm)
z = 0·175 BC − 4·17
r = 0·833
z (mm)
BC (mm)
z = 98·4 − 0·778 α
r = −0·954
z (mm)
α (degrees)

Fig. 8.9 Histogram of the measurements of angle α of hyperbolic cracks in the Assynt area.

(diagram overleaf)


n = 393
%
25
20
15
10
5
50
100
150
measurements of angle α (in degrees)

(section 6.3), are of the order of 20-30 m. In order to see whether or not friction cracks formed beneath thick ice are significantly different from those formed beneath thin ice, a comparison was made between the measured parameters of Loch Lomond Advance friction cracks and those formed beneath ice of the last ice sheet. However, no significant differences were apparent.

(b) Orientation of the friction cracks. Fig. 8.10 shows the results of the orientation analysis of 761 crescentic notches from the 12 sampling sites on the quartzite dipslope of Canisp. Several modes are apparent at each, but the major modes correlate fairly well with the direction of former ice flow as indicated by striae at the respective sampling site. At most locations there is a spread of individual orientations, which underlines the need to measure many friction cracks at a site when attempting to determine the direction of former ice flow. During field work it became apparent that the orientations of many crescentic notches were strongly influenced by structural weaknesses in the rock, the secondary crack tending to form along bedding planes, joints and cleavage planes. This might explain some of the minor modes at certain of the sites, where modal orientations occur perpendicular to the lines of weakness in the crystalline rock (Fig. 8.10).

Fig. 8.11 shows the results of the orientation analyses of the hyperbolic cracks from the same 12 sites. Once again the former ice-flow direction as indicated by local glacial striae is approximately paralleled by a major mode at each site, although other modes exist. It is interesting to note that the orientation measurements orthogonal to the line connecting the two 'horns' of a hyperbolic crack (i.e. orientations perpendicular to line BC in Fig. 8.4(b)) appear, on the whole, to give a better approximation of the former ice-flow direction.

Fig. 8.10 Rose diagrams of the orientation of crescentic notches from sampling sites on the eastern slopes of Canisp.

Key: 1. Local striae direction.

2. Trends of major lines of rock weakness.

3. Trends of minor lines of rock weakness.

(diagram overleaf)

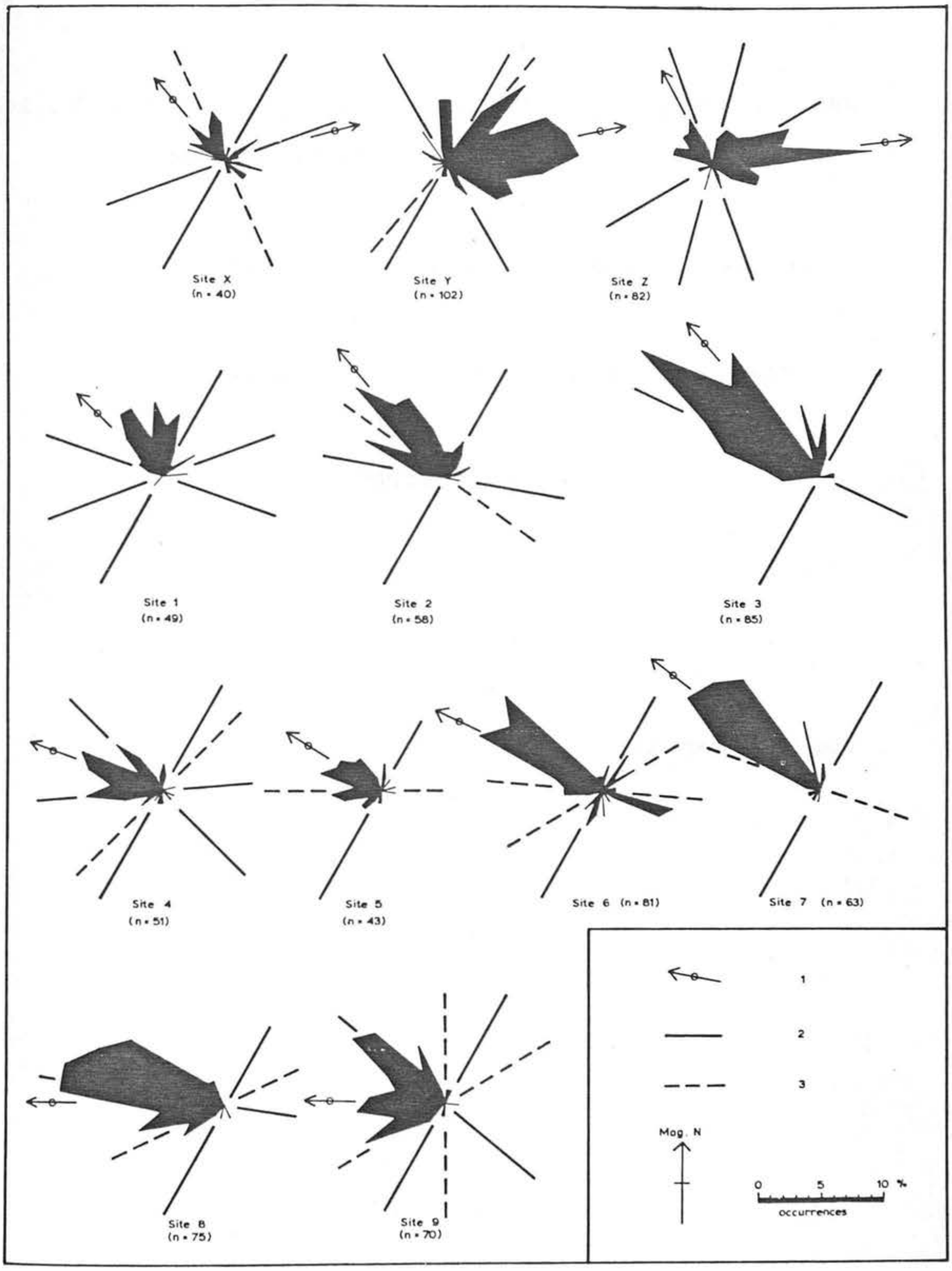

Site X
(n = 40)
Site Y
(n = 102)
Site Z
(n = 82)
Site 1
(n = 49)
Site 2
(n = 58)
Site 3
(n = 85)
Site 4
(n = 51)
Site 5
(n = 43)
Site 6 (n = 81)
Site 7 (n = 63)
Site 8
(n = 75)
Site 9
(n = 70)
1
2
3
Mag. N
0 5 10 %
occurrences

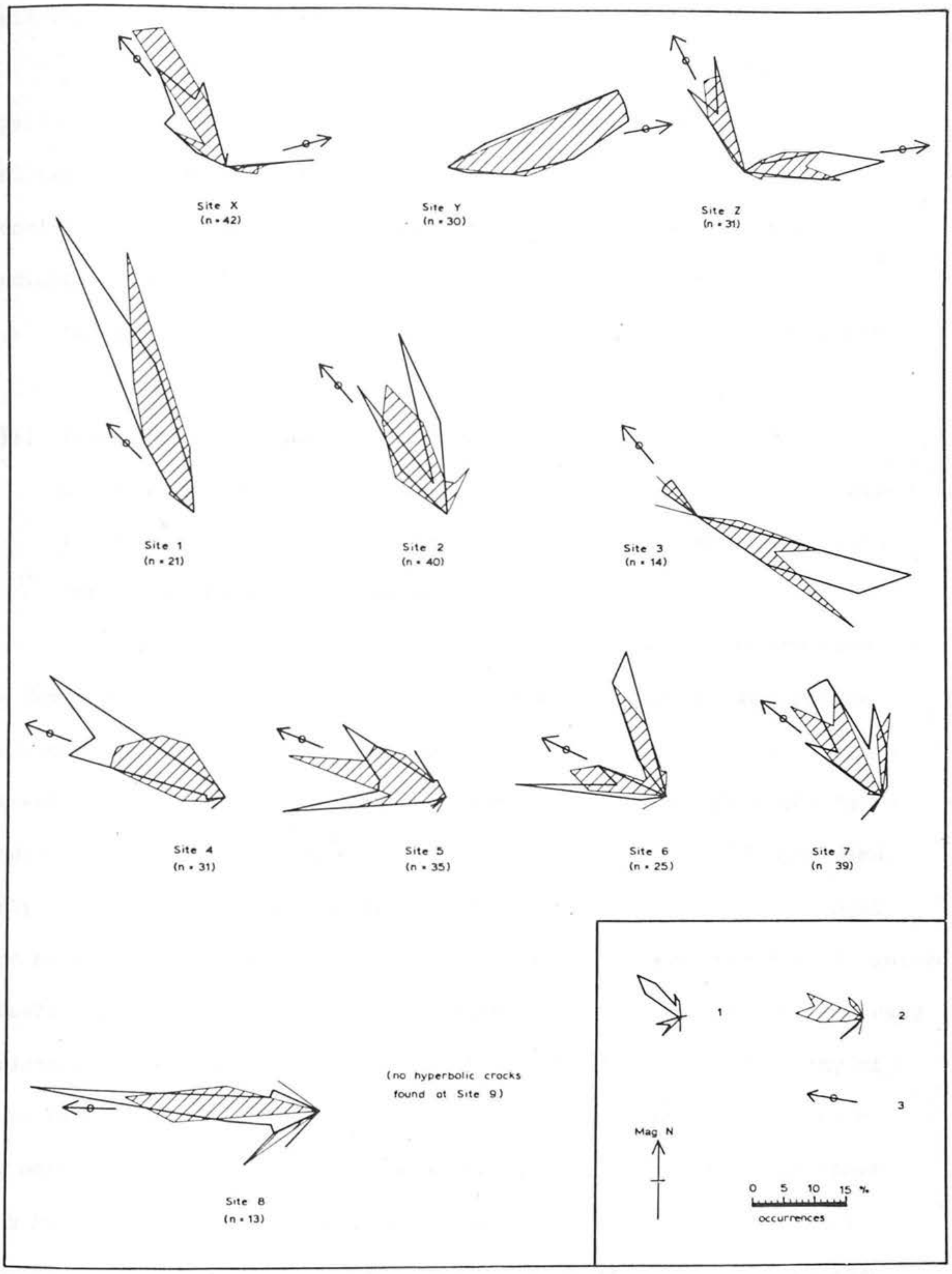
Fig. 8.11 Rose diagrams of the orientation of hyperbolic cracks from sampling sites on the eastern slopes of Canisp.

Key: 1. First orientation measurements (bisecting angle α).

2. Second orientation measurements (orthogonal to line BC).

3. Local striae direction.

(diagram overleaf)



Site X
(n = 42)
Site Y
(n = 30)
Site Z
(n = 31)
Site 1
(n = 21)
Site 2
(n = 40)
Site 3
(n = 14)
Site 4
(n = 31)
Site 5
(n = 35)
Site 6
(n = 25)
Site 7
(n 39)
Site 8
(n = 13)
(no hyperbolic cracks found at Site 9)
1
2
3
Mag N
0 5 10 15 %
occurrences

However, it must be stressed that some of the results shown in Fig. 8.11 are based on only a few hyperbolic cracks, which were not well represented at all the chosen sites.

Comparison of figures 8.10 and 8.11 reveals a fairly close similarity between the orientations of hyperbolic cracks and crescentic notches from the same site. As individuals of the two types may vary considerably from the orientation of the local striae, this may indicate one or both of the following:

(i) the effect of local ice-streaming and diversion over and around microrelief features of the bedrock surface;

(ii) that the striae and the friction cracks may not be exactly contemporaneous. The deeper friction crack features are likely to be less transitory than the striae, which often only reflect the later phases of a glaciation (chapter 4).

Modes at up to 90° from the direction of former ice flow as defined by striae at a site can be explained by local topographic influences on ice movement, but many of the sites in Fig. 8.10 possess several crescentic notches with orientations up to 180° from the former ice-flow direction; site 6 has a particularly large mode of this kind. Only site 3 in Fig. 8.11 has an enigmatic mode of this nature amongst the hyperbolic cracks. This either signifies that a hitherto unrecognised ice-flow direction exists down the quartzite dip-slope, or that reversed friction cracks can be formed orientated at 180° from the majority of such features at that place. Although it is impossible to say which is more likely to be correct, the author favours the former, as there may have been localised downslope ice movement in the Loanan valley during deglaciation of the area. Also, one might expect to find more instances of 'reversed' friction cracks if they do exist.

8.5.3 Conclusions

A rationalisation of past confused nomenclature and a classification of glacial friction cracks has been made, and four main friction crack types identified and named: crescentic notches, hyperbolic cracks, arcuate cracks and ring cracks. Quantitative analysis of over 1,300 crescentic notches and hyperbolic cracks from the Assynt area has enabled a certain amount of discussion of their size and shape, and of their value as indicators of former ice-flow directions.

Modal size of the Assynt crescentic notches is between 20 mm and 29 mm, and between 30 mm and 39 mm for the hyperbolic cracks, although the total range of sizes is much larger in both cases. The average radius of curvature for the crescentic notches is between 11 mm and 20 mm. The more complicated geometrical shape of the hyperbolic cracks proved more difficult to characterise adequately. Analysis of the size of the angle between the two 'limbs' of the crack has shown that a majority of hyperbolic cracks have an angle of between 101° and 110°.

A comparison of the orientation of the crescentic notches and hyperbolic cracks with the average orientation of glacial striae at each site has shown that friction cracks may indeed be used as directional indicators of former ice flow, although they are not orientated as consistently as the striae. The spread of orientation measurements at any particular site, which often show a number of minor modes, may in part be due to structural weaknesses within the bedrock, as well as stress-field and ice-flow differences caused by minor topographic irregularities. In order to attempt to overcome this problem, the orientation of a large number of friction cracks should be measured where possible. Measurement of the orientation perpendicular to a line joining the 'horns' of hyperbolic cracks appears to give a better estimation of the former ice-flow direction than orientations based on the angle between the 'limbs' of the cracks.

CHAPTER 9: GEOMORPHOLOGY OF THE ASSYNT KARST AREAS

9.1 INTRODUCTION

An area of approximately 34 km^2 of Cambrian dolomitic limestone crops out in the Assynt area. These calcareous rocks form some of the thickest strata of this kind in Scotland, allowing the development of a karst landscape rarely found elsewhere in the country and therefore meriting detailed study.

The petrology of the rocks has been analysed by Robertson *et al.* (1949) and Muir *et al.* (1956). The Durness limestones and dolomites, of which the Assynt dolomites are part, differ from Carboniferous limestone in several respects. They range from "fine-grained, compact porcellanous types through finely crystalline varieties to rocks in which the texture is coarsely granular" (Robertson *et al.* 1949, p. 24). The lowermost Grudaidh Formation comprises dark grey, fine-grained, distinctly granular dolomites, in most places impure : a chemical analysis of a sample of these rocks at Inchnadamph showed 52.6% $CaCO_3$, 40.2% $MgCO_3$ and 7.2% insoluble residue (*ibid.*, p. 37). The Eilean Dubh Formation, stratigraphically younger, comprises a variety of rock types, though the dominant type is a light-grey, finely crystalline dolomite, alternating with compact porcellanous varieties. These rocks have been marmorised by the Loch Borralan intrusion around Ledbeg. A chemical analysis of a sample from Elphin gave "52.40% $CaCO_3$, 38.48% $MgCO_3$ and 3.04% insoluble residue (*sic*)" (*ibid.*, p. 38). A few small outcrops of the Sailmhor Formation occur in the Assynt area, comprising fine-grained, light grey dolomites alternating with coarser varieties; no chemical analyses are reported from this group.

The Assynt dolomites crop out in the Moine Thrust belt between the Sole Thrust and the Glencoul Thrust - Ben More Thrust planes (Fig. 2.2). Geological sections through the area (Peach et al. 1907, Fig. 39, Elliott & Johnson 1980, Fig. 15) show that the substantial thickness of the dolomites in Assynt is due to repeated thrusting of the rocks to form an imbricated stack of thin slices of dolomite one on top of another. It will be shown that this peculiar geological structure has greatly influenced the hydrology and speleology of the area.

Although the existence of the Assynt caves has long been known (e.g. Lennie 1911), speleological research in the area was first undertaken by the Sheffield University Mountaineering Club in the years following the Second World War (S.U.M.C. 1950, 1953). The results of this work were summarised and considerably enlarged upon by Ford (1959). He suggested that the caves first formed under phreatic conditions (i.e. passages completely water-filled) during the Tertiary period when the dolomite rocks were covered by other strata in the Moine Thrust belt. Subsequent repeated glaciation stripped away these overlying strata to reveal the dolomite progressively. Phases of vadose activity (i.e. passages containing subterranean streams with air-spaces above) in interglacial periods resulted in downward development of caves, with glaciation possibly removing the abandoned upper parts of cave systems. Intense vadose activity was initiated by rapid runoff during the last deglaciation. At present less vadose action continues, with frequent inundation of cave passages by floods in high-water conditions. Ford described the known caves in turn, concluding that many were immature systems dominated at present by vadose action, attributing a Postglacial age to a large number of the cave passages.

Warwick (1962) included the Assynt area in his summary of British caving areas. He relied totally on Ford's paper, adding nothing new. Subsequent exploration work in the area has been dominated by the Grampian Speleological Group. Summaries of progress are to be found in the various issues of the Group's Bulletin and the speleological guidebook to the area (Jeffries 1972). Few major cave finds have been made in the area since the latter publication, except for the discovery of extensive lengths of cave passage in Uamh an Claonaite (summarised in Jeffries & Young (1980) and Young (1980)). The dolomite outcrops can be divided into five main areas (Fig. 9.1). Only one cave has been found in the Achmore area (Allt a' Chalda Mor Stream Cave at NC 25412355) although the area is pitted by numerous small dolines. Caves have yet to be discovered in the marmorised dolomite of the Ledbeg valley east of Lyne Croft and in the east of the Knockan basin. This study therefore concentrates on the Traligill basin, the Allt nan Uamh basin and the Abhainn a' Chnocain basin.

9.2 DESCRIPTIONS OF THE MAIN DRAINAGE BASINS

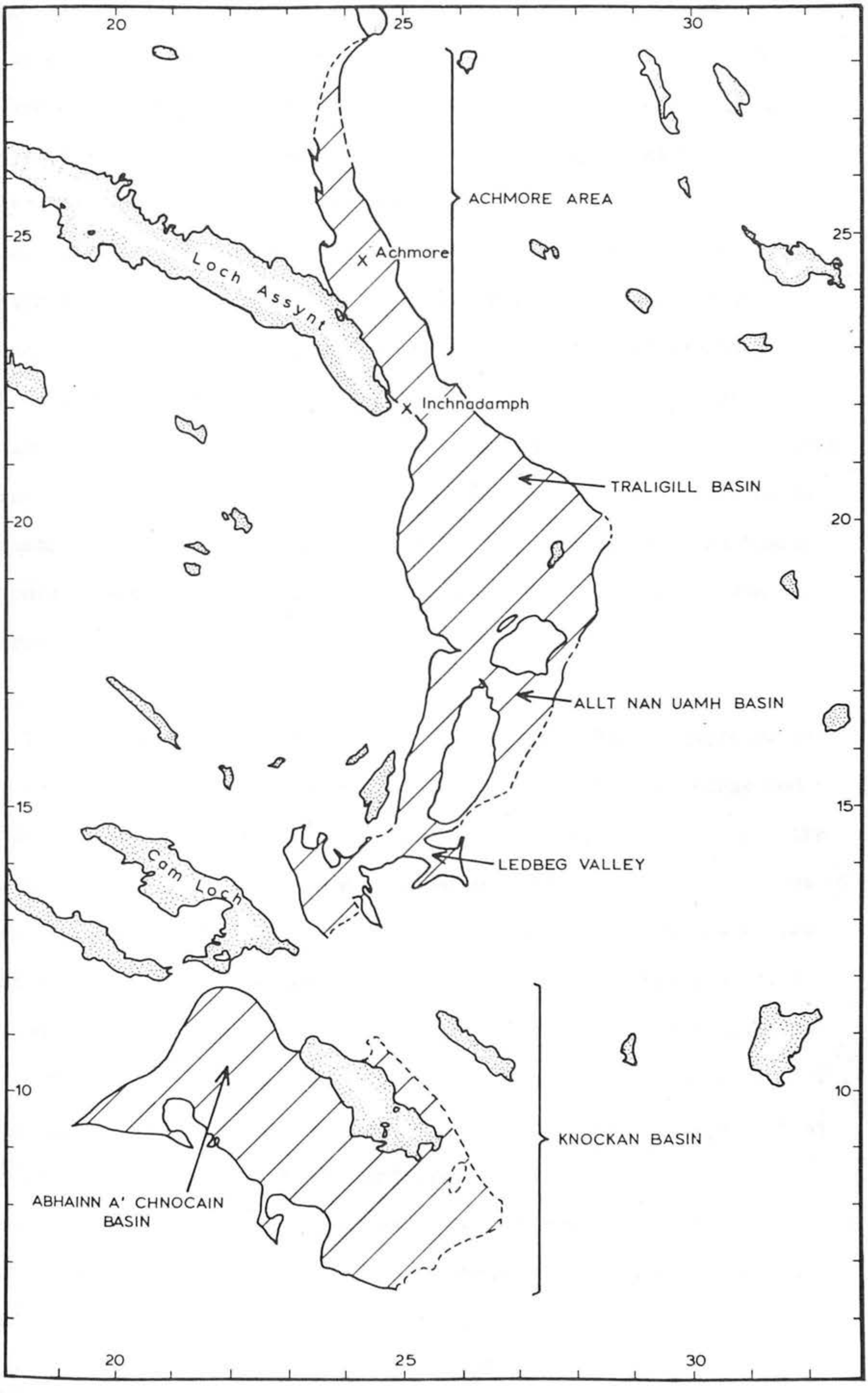
9.2.1 The Traligill Basin

(a) Major Caves

The caves of the Traligill basin are briefly described in Jeffries (1972). Six of them can be considered major caves and, because of detailed work on their geomorphology and sedimentary infill, are described below.

Fig. 9.1 The main areas of dolomite in the Assynt area.

(diagram overleaf)



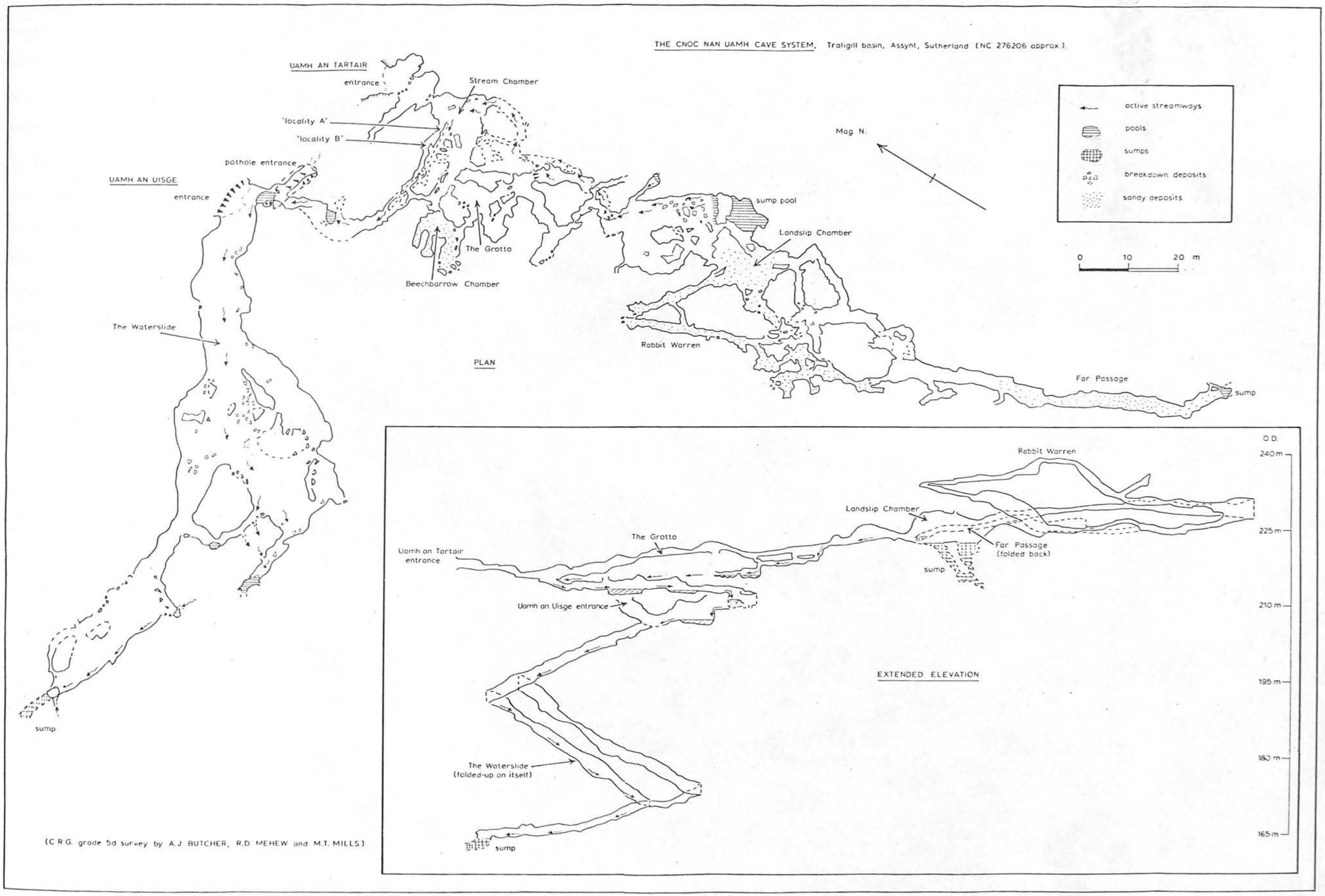
20
25
30
ACHMORE AREA
Achmore
Loch Assynt
Inchnadamph
TRALIGILL BASIN
ALLT NAN UAMH BASIN
LEDBEG VALLEY
Cam Loch
KNOCKAN BASIN
ABHAINN A' CHNOCAIN
BASIN
25
20
15
10

(i) Uamh Cailliche Peireag (NC 27652033). This cave as presently known is approximately 40 - 50 m long, and with clastic sediments at both ends is presumably part of a larger cave system. The existing passages are phreatic tubes that lie just above an impermeable igneous intrusion which is well exposed in the dry stream bed at the cave entrance. Although the walls have been extensively shattered, a few remaining scallop features indicate they were formed by phreatic water flowing from south to north. The cave passages have been intersected by a waterfall retreating along the Allt a' Bhealaich. The sedimentary infill mainly comprises a coarse cobble and gravel mixture with a sandy matrix, though patches of grey silt and red-brown gravelly sands were also noted. Many sub-angular to rounded erratic clasts were identified. The cobbles and gravels have been partially eroded at some stage, followed by the deposition of a thin flowstone layer.

(ii) Cnoc nan Uamh system (approx. NC 276206). This integrated cave system with its three entrances (Uamh an Tartair, Uamh an Uisge and a large pothole) is one of the largest caves in the Assynt area. The cave survey (Fig. 9.2) shows a complex upper phreatic series of cave passages, with the present streamway following a lower vadose series after emerging from the sump in Landslip Chamber, turning finally to cascade down a thrust plane (referred to hereafter as the Traligill Main Thrust) in Uamh an Uisge (The Waterslide). The phreatic form of the Rabbit Warren passages indicates an early water-table position at or above 240 m O.D. In flood conditions, formerly-phreatic Far Passage acts as a streamway, as shown by the presence of vegetation debris, peaty mud and areas of clean pebble floors. The completely inactive Rabbit Warren is almost totally filled in places with pale

Fig. 9.2 Survey of the Cnoc nan Uamh cave system.

(diagram overleaf)



THE CNOC NAN UAMH CAVE SYSTEM, Traligill basin, Assynt, Sutherland (NC 276206 approx.).
UAMH AN TARTAIR
entrance
Stream Chamber
'locality A'
'locality B'
pothole entrance
UAMH AN UISGE
entrance
The Grotto
Beechbarrow Chamber
sump pool
Landslip Chamber
Mag. N.
active streamways
pools
sumps
breakdown deposits
sandy deposits
0
10
20 m
The Waterslide
PLAN
Rabbit Warren
Far Passage
sump
sump
O.D.
240 m
225 m
210 m
195 m
180 m
165 m
Rabbit Warren
Landslip Chamber
The Grotto
Uamh an Tartair entrance
Far Passage (folded back)
sump
Uamh an Uisge entrance
EXTENDED ELEVATION
The Waterslide (folded-up on itself)
sump
(C.R.G. grade 5d survey by A.J. BUTCHER, R.D. MEHEW and M.T. MILLS)

yellow laminated silts and sands. Breakdown has drastically modified the original phreatic passages elsewhere in the cave.

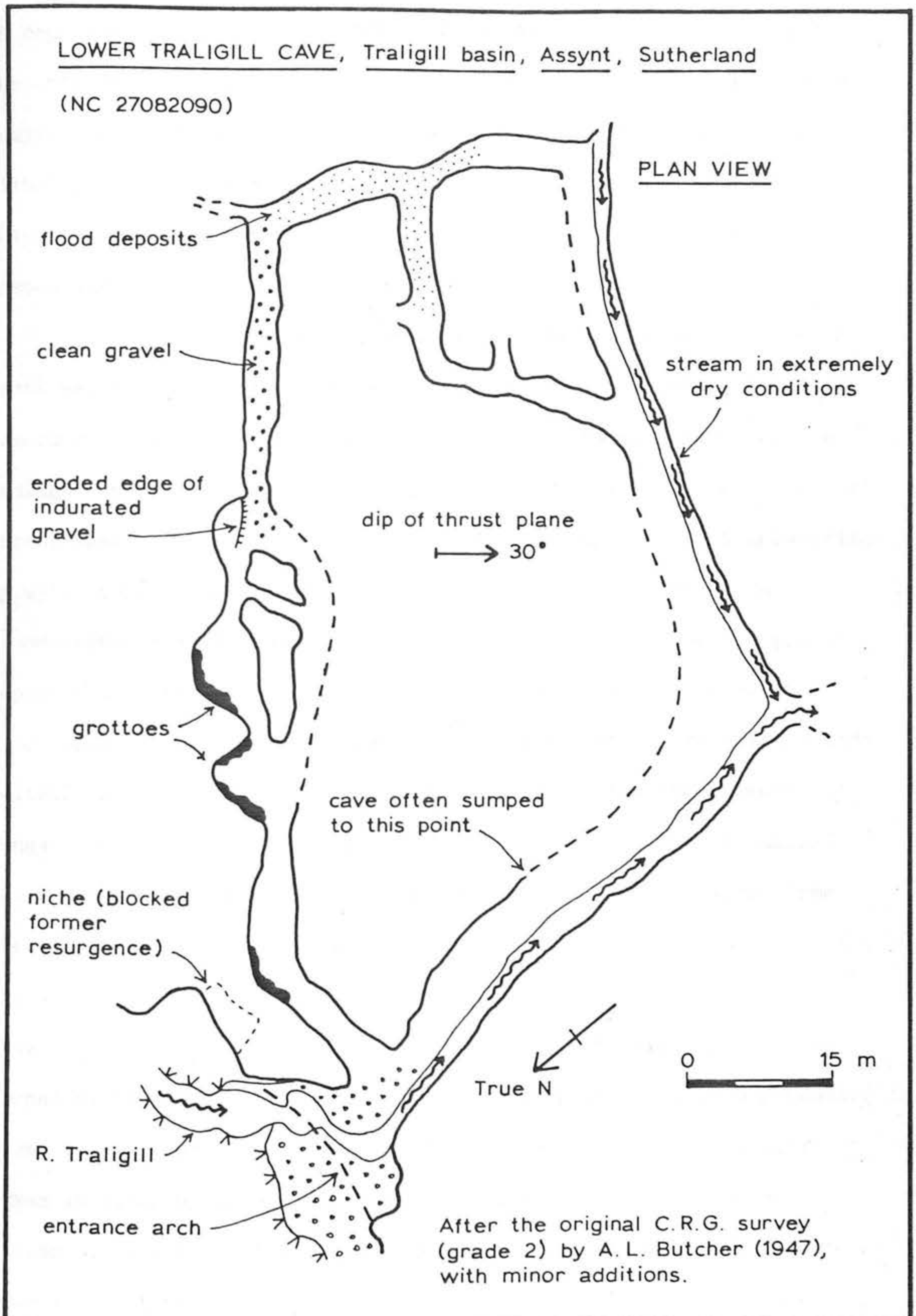
The general stratigraphic sequence of clastic fill in the caves is a composite one, there being no one locality where the complete stratigraphy of over 8 m of sediments can be seen. However, two separate sections, exposed in the high-level rift leading off to the west from the Stream Chamber (A and B in Fig. 9.2), can be combined to show representatives of all the strata. Pale yellow laminated silts and sands are separated by a sharp contact from overlying fluvial gravels and cobbles with a high erratics content, forming a channel fill, spreading laterally into a red-brown sand. This passes upwards into sand and silt lenses. The whole sedimentary sequence is topped by breakdown products and stalagmite deposition elsewhere in the cave.

The Waterslide in Uamh an Uisge is clean-washed. On the periphery of the active streamway, the presence of gravels, some indurated with calcite, suggests that The Waterslide was possibly once choked with coarse sediments that may have precluded its use by the stream, which would have been forced to resurge at this point. Indeed, an entrance blocked with breakdown debris and thin flowstone exists at the top of the thrust plane, leading into a series of dry channels outside the cave. The present entrances to Uamh an Uisge and the nearby pothole were probably formed by roof collapse or by glacial erosion. A dry channel also leads from the entrance of Uamh an Tartair, higher up the hill.

(iii) Lower Traligill Cave (NC 27092090). This cave has developed on two levels along the strike of the Traligill Main Thrust plane (Fig. 9.3). The lower of the two levels is an active streamway, being frequently sumped and accessible only in extremely dry conditions.

Fig. 9.3 Survey of Lower Traligill Cave.

(diagram overleaf)


LOWER TRALIGILL CAVE, Traligill basin, Assynt, Sutherland
(NC 27082090)
PLAN VIEW
flood deposits
clean gravel
eroded edge of indurated gravel
grottoes
niche (blocked former resurgence)
R. Traligill
entrance arch
dip of thrust plane
30°
stream in extremely dry conditions
cave often sumped to this point
True N
0
15 m
After the original C.R.G. survey (grade 2) by A.L. Butcher (1947), with minor additions.

The upper level is dry except in flood conditions. This upper level is represented at the cave entrance by a niche on the northern side of the entrance archway. Incision by the Traligill River has exposed a section, comprising up to 1 m of angular gravel containing erratic lithologies in a red silty matrix, overlain by 30 cm of angular, dolomitic coarse gravel cemented by porous travertine and topped by moss-covered flowstone.

Within the upper levels of the cave, deposits comprise a basal water-worn gravel of erratic lithologies mixed with sub-angular breakdown fragments, cemented and largely covered with flowstone and stalagmites of brown calcite. That this brown calcite is not the most recent phase of speleothem deposition is shown by its partial covering by white calcite associated with modern roof-drip. Present-day floodwaters have excavated the lowest levels of the indurated gravel deposits to reveal the smooth, inclined bedrock floor. Above the flood level, the cave roof possesses many small straw-stalactites and helictites, the latter thought to be associated with the presence of fungal hyphae (Ford 1959). In the lower levels of the cave, walls, floor and roof are covered in a dark brown silty mud deposited from receding waters after flooding.

(iv) <u>Tree Hole Cave</u> (NC 26952097). The present large chamber was formed by breakdown of the cave roof into passages excavated phreatically along the Traligill Main Thrust. Evidence of the initial phreatic phase is given by solution pockets in the roof near the present entrance. A stream flows beneath the large breakdown blocks, lower down the inclined thrust plane.

(v) <u>Glenbain Hole</u> (NC 2650-2169). A large entrance in an open shakehole is partly blocked by breakdown deposits. Flecks of charcoal incorporated in muddy slump deposits in the entrance chamber yielded a radiocarbon date of 2,560 $\pm$ 525 (S-158) (Walker 1973). The entrance chamber is floored with breakdown, and farther back narrows to a tight rift followed by a trickle of water. The cave passage continues to descend what is probably a thrust plane until a sump prohibits further penetration.

(vi) Firehose Cave (NC 26352160). This constricted phreatic tube, carrying an active stream, has formed along a thrust plane which may be the Traligill Main Thrust (Fig. 9.4). There is a decorated chamber at the back of the cave, which was probably connected at one stage to the Glenbain Hole system.

(b) Surface features and hydrology

The Traligill basin drainage begins on the quartzite slopes of Breabag, Beinn an Fhurain and Conival. Waters from the first collect in the upper basin, known as Cuil Dubh, whence they flow to a sink in a blind valley (Cuil Dubh Sink) soon after passing from the quartzite onto the dolomite area (Fig. 9.5). From here the underground drainage flows northwards in fissures and passages too constricted to follow. On the north bank of the stream just prior to Cuil Dubh Sink, a series of degraded channels leads to two large enclosed depressions, c. 10 m deep in places, separated by an igneous intrusion. Vegetation debris close to the top of these depressions shows that they take drainage waters in times of flood, presumably when the discharge of the stream from Cuil Dubh exceeds the capacity of the sink. Sections in the walls of the enclosed depressions and channels show that these features are cut in rock and covered with a veneer of till that is

Fig. 9.4 Survey of Firehose Cave.

(diagram overleaf)

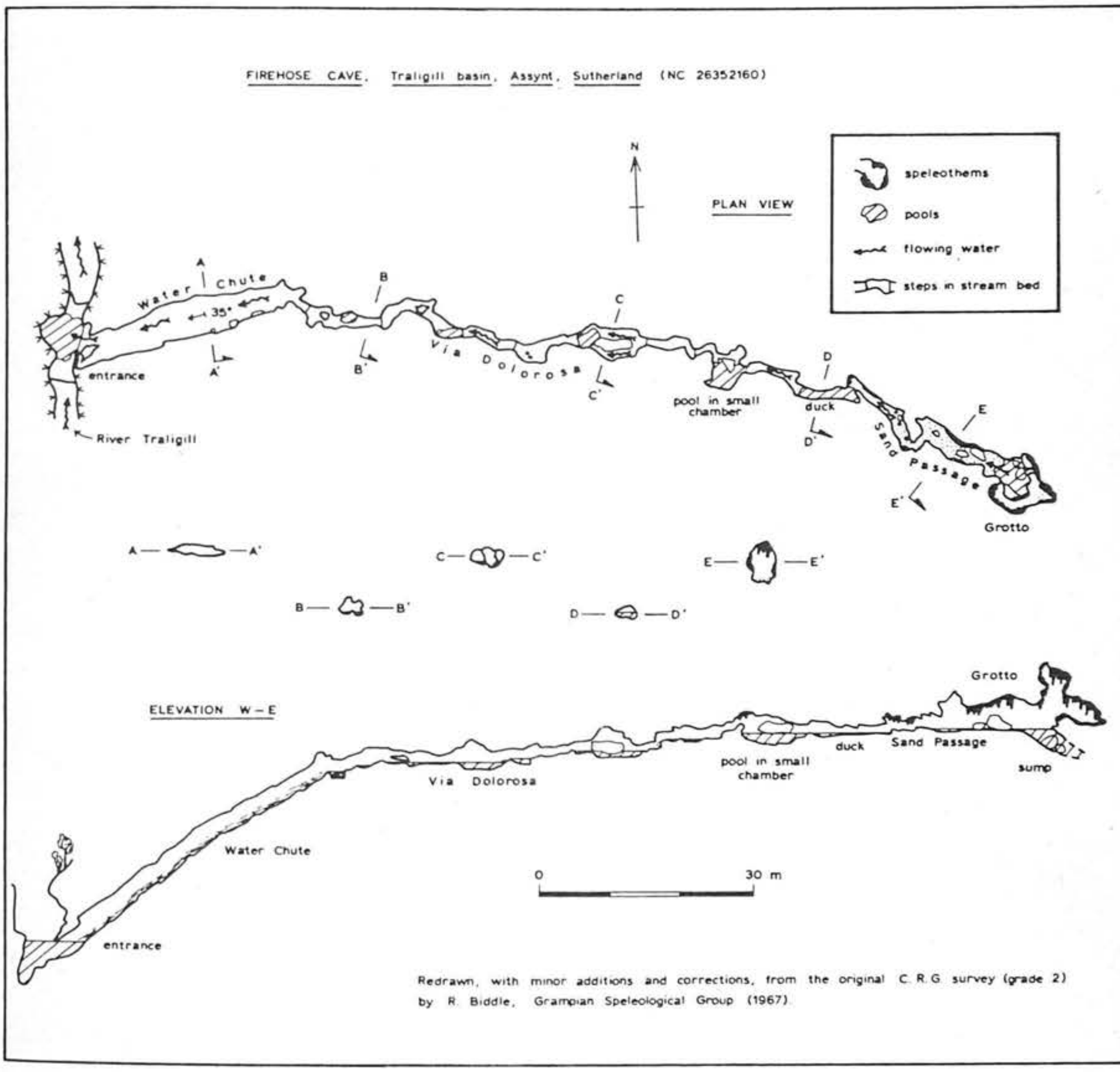
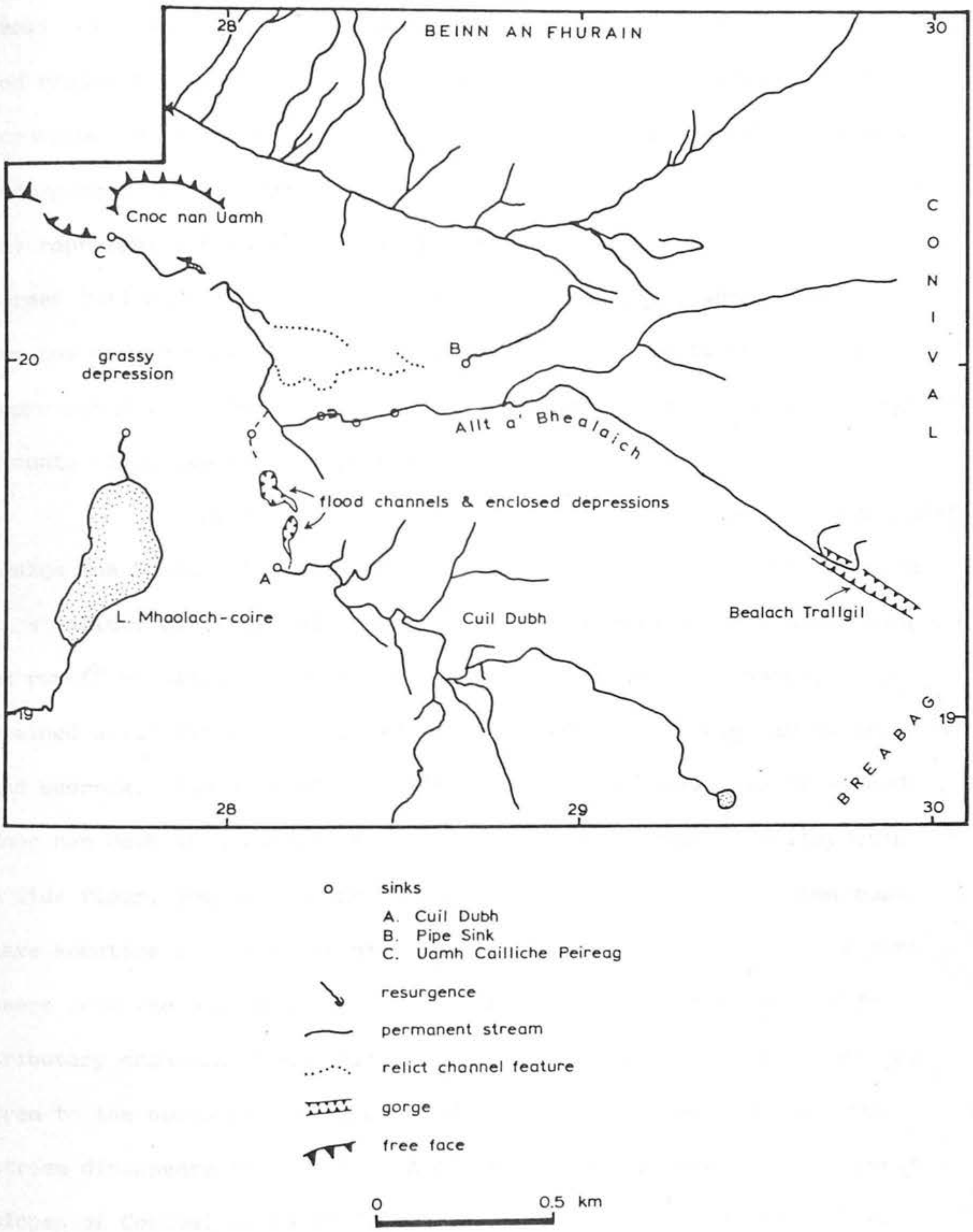

FIREHOSE CAVE, Traligill basin, Assynt, Sutherland (NC 26352160)
N
PLAN VIEW
speleothems
pools
flowing water
steps in stream bed
Water Chute
35°
entrance
River Traligill
Via Dolorosa
pool in small chamber
duck
Sand Passage
Grotto
A
A'
B
B'
C
C'
D
D'
E
E'
ELEVATION W - E
Grotto
duck
Sand Passage
sump
pool in small chamber
Via Dolorosa
Water Chute
entrance
0
30 m
Redrawn, with minor additions and corrections, from the original C.R.G. survey (grade 2)
by R. Biddle, Grampian Speleological Group (1967).

Fig. 9.5 The upper reaches of the Traligill drainage basin.

(diagram overleaf)


BEINN AN FHURAIN
28
30
Cnoc nan Uamh
C
CONIVAL
B
20
grassy depression
Allt a' Bhealaich
flood channels & enclosed depressions
A
L. Mhaolach-coire
Cuil Dubh
Bealach Trallgil
19
BREABAG
28
29
30
sinks
A. Cuil Dubh
B. Pipe Sink
C. Uamh Cailliche Peireag
resurgence
permanent stream
relict channel feature
gorge
free face
0
0.5 km

mostly vegetated. The depressions are orientated along the outcrop of a minor thrust fault. Although not topographically continuous, small collapses in the peat, apparently leading into the bedrock, and restricted gravel spreads leading from bouldery sections (both occurring in the bottom of each depression), may represent sinks and resurgences respectively in high-water conditions. The two depressions may represent successive fossil blind valleys acting as sinks for former Cuil Dubh streams, approaching from the east, whose channels are now hidden beneath the blanket peat. It is likely that these depressions were active sinks during the last deglaciation when large amounts of meltwater were present.

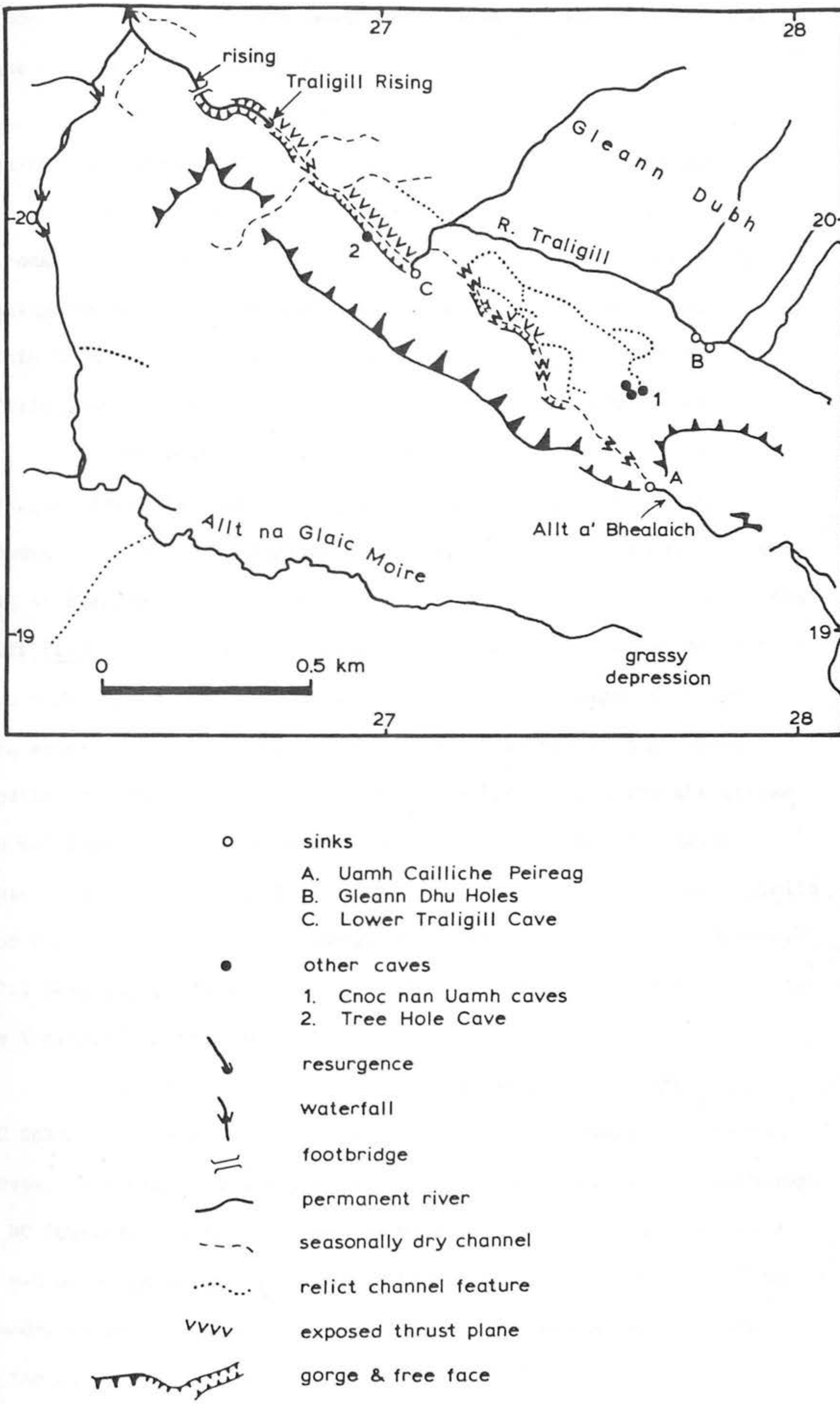
To the north of the second depression, the Allt a' Bhealaich drains the Bealach Trallgil and Conival (Fig. 9.5). This stream sinks at a variety of places after reaching the dolomite bedrock, depending on runoff conditions. Sections in its banks reveal laminated, fine-grained alluvium (e.g. at NC 28101991), coarse till (e.g. at NC 28152000) and bedrock. The Allt a' Bhealaich continues north-westwards towards Cnoc nan Uamh in a narrow channel that follows a shallow valley with a wide floor, suggesting that the surface drainage in this area must have sometime been much greater (e.g. during deglaciation). The burn seeps into the peat before reaching Cnoc nan Uamh. A system of dry tributary channels, much infilled by blanket peat, extends across the area to the north of the Allt a' Bhealaich and joins it before the stream disappears (Fig. 9.5). A small stream initiated on the lowest slopes of Conival sinks at Pipe Sink (NC 28672000). To the west of Cuil Dubh Sink, Loch Mhaolach-coire is drained by a small stream which sinks amongst boulders at NC 27691978. This drainage course is continued northwards by a dry channel cut in the peat and till, leading into a grassy, enclosed area which is the source of the Allt na Glaic Moire.

The Allt a' Bhealaich re-emerges from the peat in the middle section of the Traligill basin (Fig. 9.6), though it is little more than a trickle of water as much of the drainage has gone underground before this point is reached. It formerly used a waterfall at NC 27652033, but progressive retreat of the fall has intersected part of a cave system (Uamh Cailliche Peireag) which acts as a flood resurgence. The Allt a' Bhealaich continues from here as a dry channel except in conditions of high discharge. A number of dykes cross its course, forming dry waterfalls. In its lower reaches the dry stream bed follows the outcrop of a thrust plane, in the angle between a small bluff (forming the edge of the upper thrust mass) and the exposed surface of the inclined thrust plane which forms the northern bank. The Allt a' Bhealaich is eventually confluent with the River Traligill.

Gleann Dubh carries the Traligill, which drains the SW slopes of Beinn an Fhurain. Gleann Dubh is in places deeply incised in rock and drift though possessing a wide bottom, which suggests that formerly the valley carried a more powerful river. Gleann Dubh is strongly fault-controlled, its western end following the Traligill Pass cross fault (Elliott & Johnson 1980, Fig. 15) which controls the orientation of the Bealach Trallgil. At approximately NC 277207, tight fissures on the south bank of the Traligill (Gleann Dhu Holes) take a proportion of the river underground so that in very dry weather the river bed can be dry for c. 300 m downstream (Ford 1959). At NC 272210 the Traligill turns abruptly southwards, cutting through a ridge (i.e. down the dip of the Traligill Main Thrust: Fig. 9.6), sinking at Lower Traligill Cave and other sinks nearby. A continuation of Gleann Dubh can be followed north-westwards for a further 200 m

Fig. 9.6 The middle reaches of the Traligill drainage basin.

(diagram overleaf)



27
28
rising
Traligill Rising
Gleann Dubh
R. Traligill
20
2
C
B
1
A
Allt a' Bhealaich
Allt na Glaic Moire
19
0
0.5 km
grassy depression
27
28
sinks
A. Uamh Cailliche Peireag
B. Gleann Dhu Holes
C. Lower Traligill Cave
other caves
1. Cnoc nan Uamh caves
2. Tree Hole Cave
resurgence
waterfall
footbridge
permanent river
seasonally dry channel
relict channel feature
exposed thrust plane
gorge & free face

before this, too, turns south-westwards down the dip of the thrust plane to rejoin the Traligill.

Between the western ends of Gleann Dubh and the Allt a' Bhealaich lies Cnoc nan Uamh. This hill contains the integrated Cnoc nan Uamh cave system. A complex network of dry channels, cut in rock, leads from Uamh an Tartair and Uamh an Uisge, eventually joining the Allt a' Bhealaich (Fig. 9.6). These channels must relate to a time when water resurged from these caves. Peat partly infills some of the channels, indicating that they are relict features.

The Traligill valley extends north-westwards from Lower Traligill Cave, following the Traligill Main Thrust in the angle between cliff and thrust plane surface, thus resembling parts of the Allt a' Bhealaich. Arches and other solutional features occur on the bluff (i.e. the SW bank), providing evidence of the former position of cave passages or the river at a higher level, now exposed by erosion. This section of the Traligill valley is dry except in high water conditions. The present stream flows parallel to the surface stream bed but farther down the thrust plane; access to this stream is achieved in Tree Hole Cave. The exposed thrust plane surface exhibits good karren features : both 'Rundkarren' and 'Rillenkarren' (Jennings 1971; Sweeting 1972) are present. Several seasonally dry rivers join the Traligill in this area (Fig. 9.6).

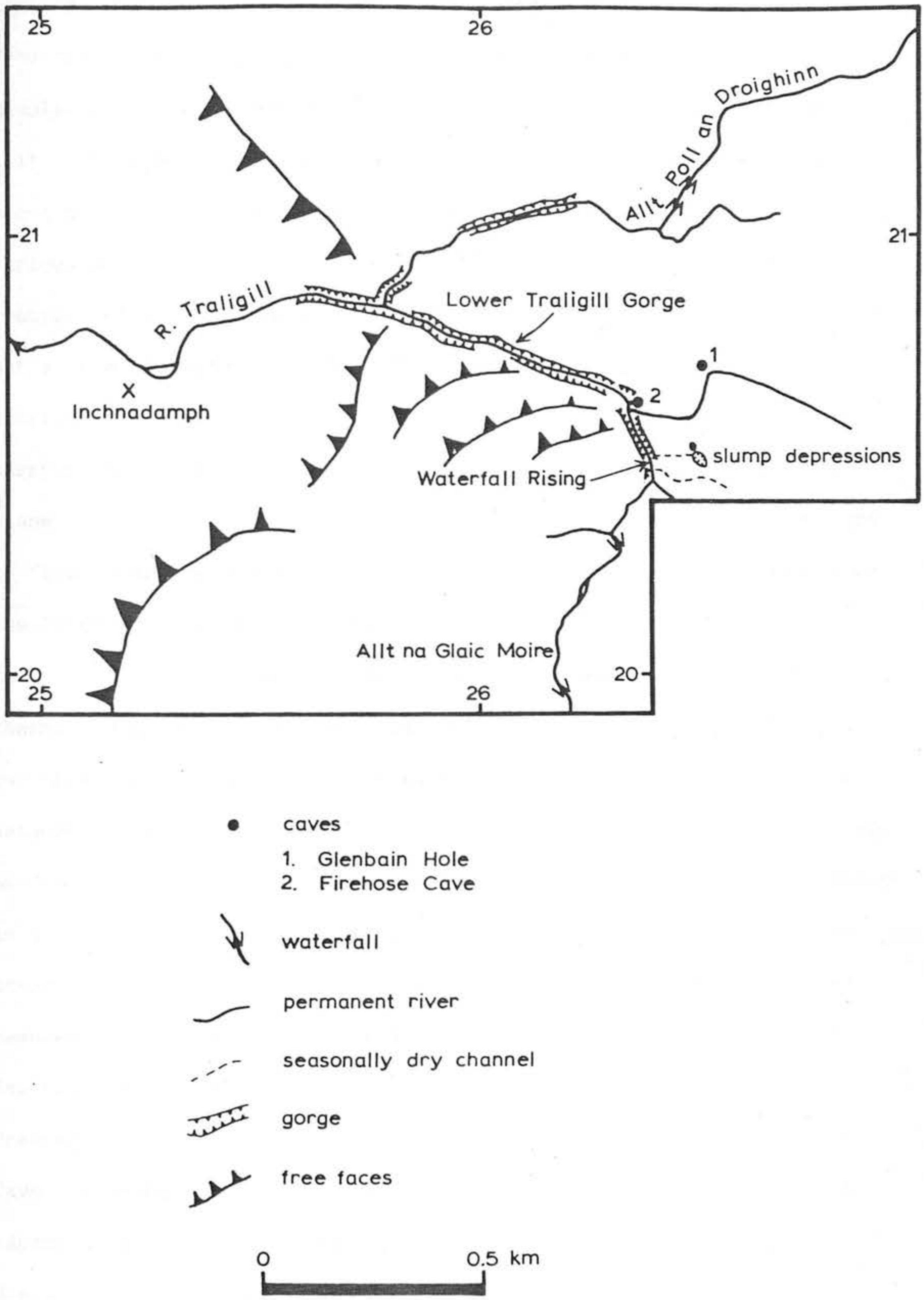
The Traligill River re-emerges at Traligill Rising (NC 26732123); from this point the river channel always carries a stream. A further resurgence occurs just downstream of the footbridge at NC 26552132. Between these two risings the Traligill occupies a 10 m-deep gorge which possibly was formed by headward erosion of the now-dry waterfall above Traligill Rising. Downstream of the gorge, as far as its confluence with the Allt na Glaic Moire (Fig. 9.6), the

Traligill occupies a comparatively broad channel. The Allt na Glaic Moire, which originates in the grassy depression north of Loch Mhaolach-coire, flows westwards along the outcrop of a thrust fault, then turns northwards to cut across the local geological structure in a series of waterfalls before joining the Traligill River.

From this confluence, the lower reaches of the Traligill River (Fig. 9.7) flow in a narrow gorge (hereafter called the Lower Traligill Gorge), at the head of which a resurgence at NC 26392148 in the east bank (Waterfall Rising) adds to the discharge of the river. The existence of the gorge at this point relates to geological structure : from the Allt na Glaic Moire confluence, the Traligill cuts across the strike of the successive thrust masses of the Fiadhag imbricate stack (Elliott & Johnson 1980, Fig. 15) whereas previously the river tended to flow sub-parallel to the thrust plane outcrops. Approximately 150 m downstream from Waterfall Rising, a powerful resurgence occurs from Firehose Cave. Approximately 60 m higher up the hillside above the resurgence, a small stream running off the quartzites by-passes the large entrance of Glenbain Hole where it doubtless once went underground, and flows down the hillside through structural ridges to join the Traligill near Firehose Cave. Several seasonally dry channels occur nearby (Fig. 9.7), one running from depressions caused by slumps in the local till. The Lower Traligill Gorge eventually gives way to the alluvial flats on which part of Inchnadamph stands. A major right-bank tributary, the Allt Poll an Droighinn, flows through a small gorge prior to joining the Traligill near the end of the Lower Traligill Gorge. Excavation along a thrust plane at the lower end of the Allt Poll an Droighinn has resulted in the formation of two small caves, Begba Hole and Uamh nan Calumen (NC 257218 and NC 258219 respectively).

Fig. 9.7 The lower reaches of the Traligill drainage basin.

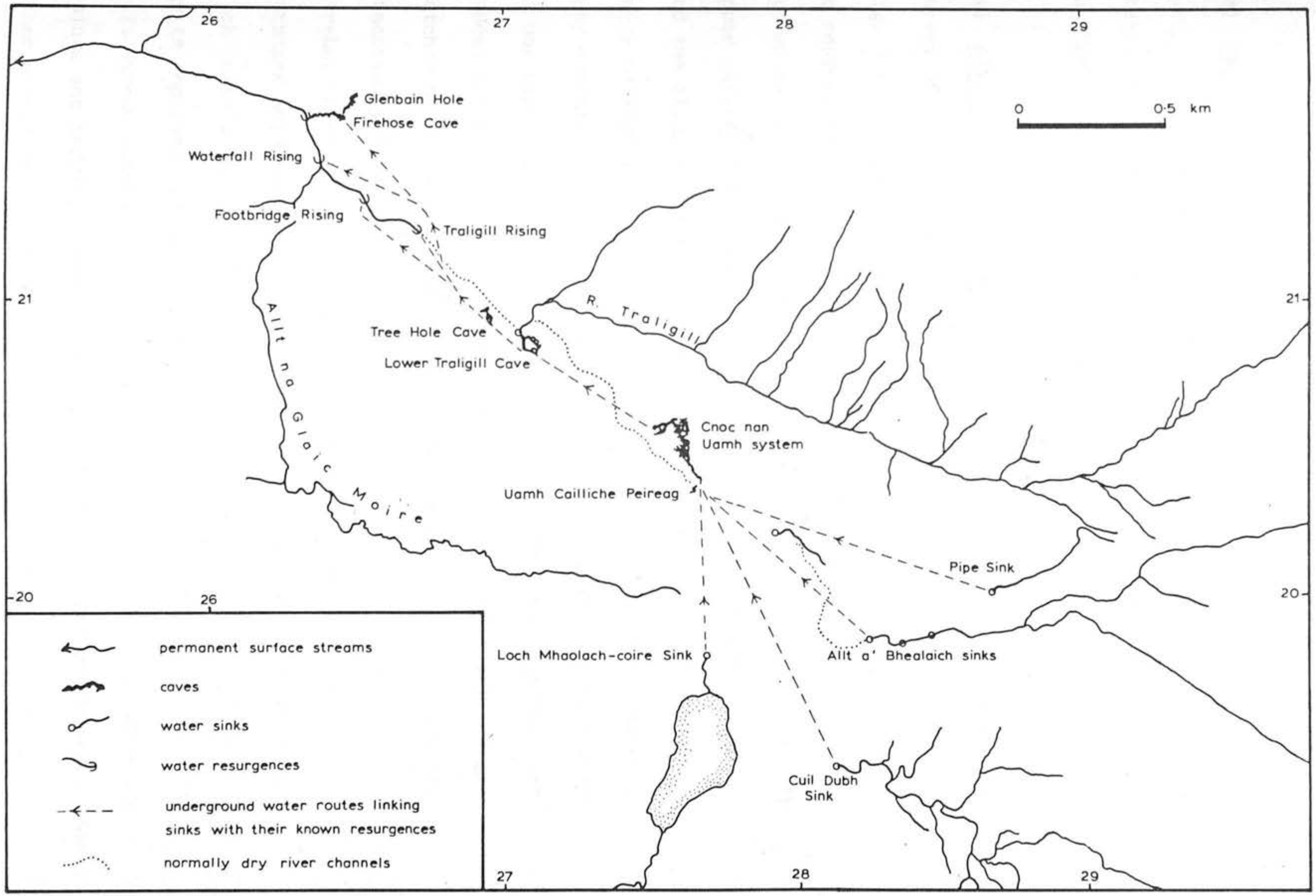
(diagram overleaf)


25
26
21
21
20
20
25
26
Allt Poll an Droighinn
R. Traligill
Lower Traligill Gorge
1
2
X
Inchnadamph
slump depressions
Waterfall Rising
Allt na Glaic Moire
caves
1. Glenbain Hole
2. Firehose Cave
waterfall
permanent river
seasonally dry channel
gorge
free faces
0
0.5 km

The hydrology of the basin has been investigated by dye-tracing techniques (P. L. Smart, pers. comm.) (Fig. 9.8). All water from the upper basin (i.e. from Pipe Sink, Cuil Dubh Sink, Loch Mhaolach-coire Sink and the various sinks along the course of the Allt a' Bhealaich) drains through the Cnoc nan Uamh cave system and along the Traligill Main Thrust plane. It is joined by water entering various sinks around Lower Traligill Cave. The combined waters resurge not only at Traligill Rising and the rising near the footbridge, but also at Waterfall Rising and Firehose Cave. The latter also carries the water from Glenbain Hole. When the discharge exceeds the carrying capacity of the cave system along the Traligill Main Thrust plane, the flow at Lower Traligill Cave is reversed and water resurges to flow along the normally dry river bed. In times of high discharge, the lower (dry) part of the Allt a' Bhealaich channel carries water that resurges from Uamh Cailliche Peireag. Under flood conditions the channels and depressions near Cuil Dubh Sink probably fill with water, but this has not been witnessed by the writer. The presence of the network of dry channels leading from Uamh an Tartair and Uamh an Uisge on Cnoc nan Uamh indicates that water used to resurge here either prior to the opening-up of the Traligill Main Thrust plane down which the stream currently flows at The Waterslide, or while this feature was temporarily blocked. The overall picture is thus one of progressive lowering of the regional water-table with the main subterranean drainage following the Traligill Main Thrust plane. Lower Traligill Cave was probably once a major resurgence, but Traligill Rising has captured these waters (except in times of high discharge). Traligill Rising in turn will be usurped by Waterfall Rising and Firehose Cave sometime in the future.

Fig. 9.8 Hydrology of the Traligill basin.

(diagram overleaf)


26
27
28
29
21
20
0
0·5 km
Glenbain Hole
Firehose Cave
Waterfall Rising
Footbridge Rising
Traligill Rising
R. Traligill
Tree Hole Cave
Lower Traligill Cave
Allt na Glaic Moire
Cnoc nan
Uamh system
Uamh Cailliche Peireag
Pipe Sink
Loch Mhaolach-coire Sink
Allt a' Bhealaich sinks
Cuil Dubh
Sink
permanent surface streams
caves
water sinks
water resurgences
underground water routes linking
sinks with their known resurgences
normally dry river channels

9.2.2 The Allt nan Uamh basin

(a) Major Caves

There are three major cave systems in the Allt nan Uamh basin: Allt nan Uamh Stream Cave, Uamh an Claonaite and the Creag nan Uamh caves (Jeffries 1972).

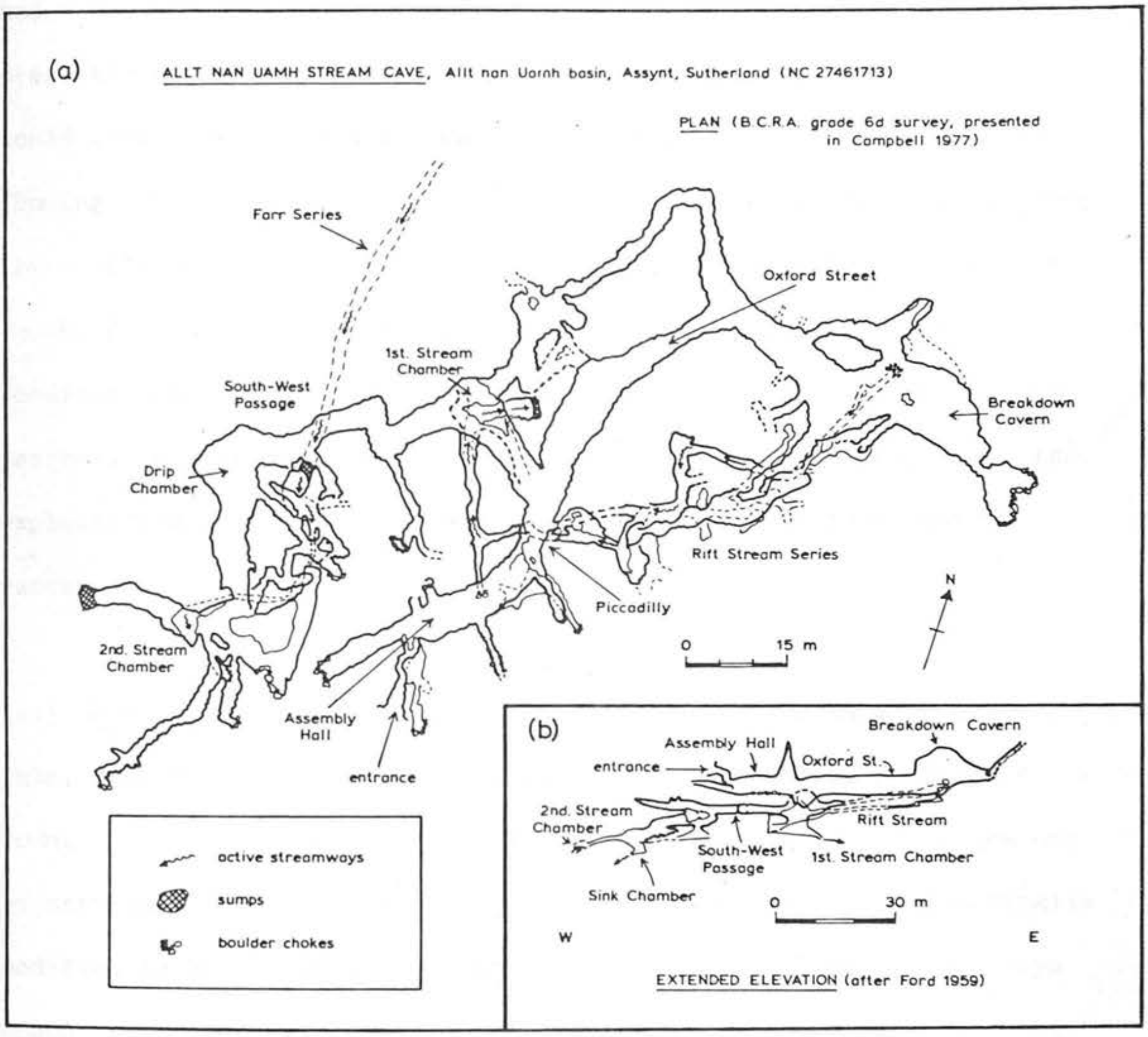
(i) Allt nan Uamh Stream Cave (NC 27461713). The most accurate survey of this complex cave (Campbell 1977) is reproduced here as Fig. 9.9(a). The only available elevation of the cave (Ford 1959) is reproduced as Fig. 9.9(b). The elevation shows that the cave system has developed on three, possibly four, separate levels. The upper series (i.e. Assembly Hall, Oxford Street, Breakdown Cavern, and the slightly lower South-West Passage and Drip Chamber) are roomy passages, phreatic in origin but much changed by breakdown. They contain large quantities of pale yellow silts and sands which at one time filled the chambers to their roofs, though subsequent vadose conditions have led to their partial excavation, leaving patches of these fine deposits on high wall-ledges. The cave is essentially free of any major speleothem formations. Active streamways occur at lower levels in the cave, in narrow vadose passages which are strongly influenced by lines of weakness in the rock (mainly inclined bedding planes according to Ford (1959)). Two quite separate streams are present in the cave (Fig. 9.9(a)). The Rift Stream enters the cave below Breakdown Chamber, flowing along joints and bedding planes to the sump in the 1st Stream Chamber. The other stream enters the cave by way of the Farr Series, flowing through the 2nd Stream Chamber to sink in Sink Chamber. The Rift Stream once occupied a higher series of rift passages, now abandoned,

Fig. 9.9 Survey of Allt nan Uamh Stream Cave.

(a) Plan (from Campbell (1977)).

(b) (inset) Extended elevation (from Ford (1959)).

(diagram overleaf)


(a)
ALLT NAN UAMH STREAM CAVE, Allt nan Uamh basin, Assynt, Sutherland (NC 27461713)
PLAN (B.C.R.A. grade 6d survey, presented in Campbell 1977)
Farr Series
Oxford Street
1st. Stream Chamber
South-West Passage
Drip Chamber
Breakdown Cavern
Rift Stream Series
Piccadilly
2nd. Stream Chamber
Assembly Hall
entrance
N
0 15 m
active streamways
sumps
boulder chokes
(b)
Breakdown Cavern
Assembly Hall
entrance
Oxford St.
2nd. Stream Chamber
Rift Stream
South-West Passage
1st. Stream Chamber
Sink Chamber
0 30 m
W
E
EXTENDED ELEVATION (after Ford 1959)

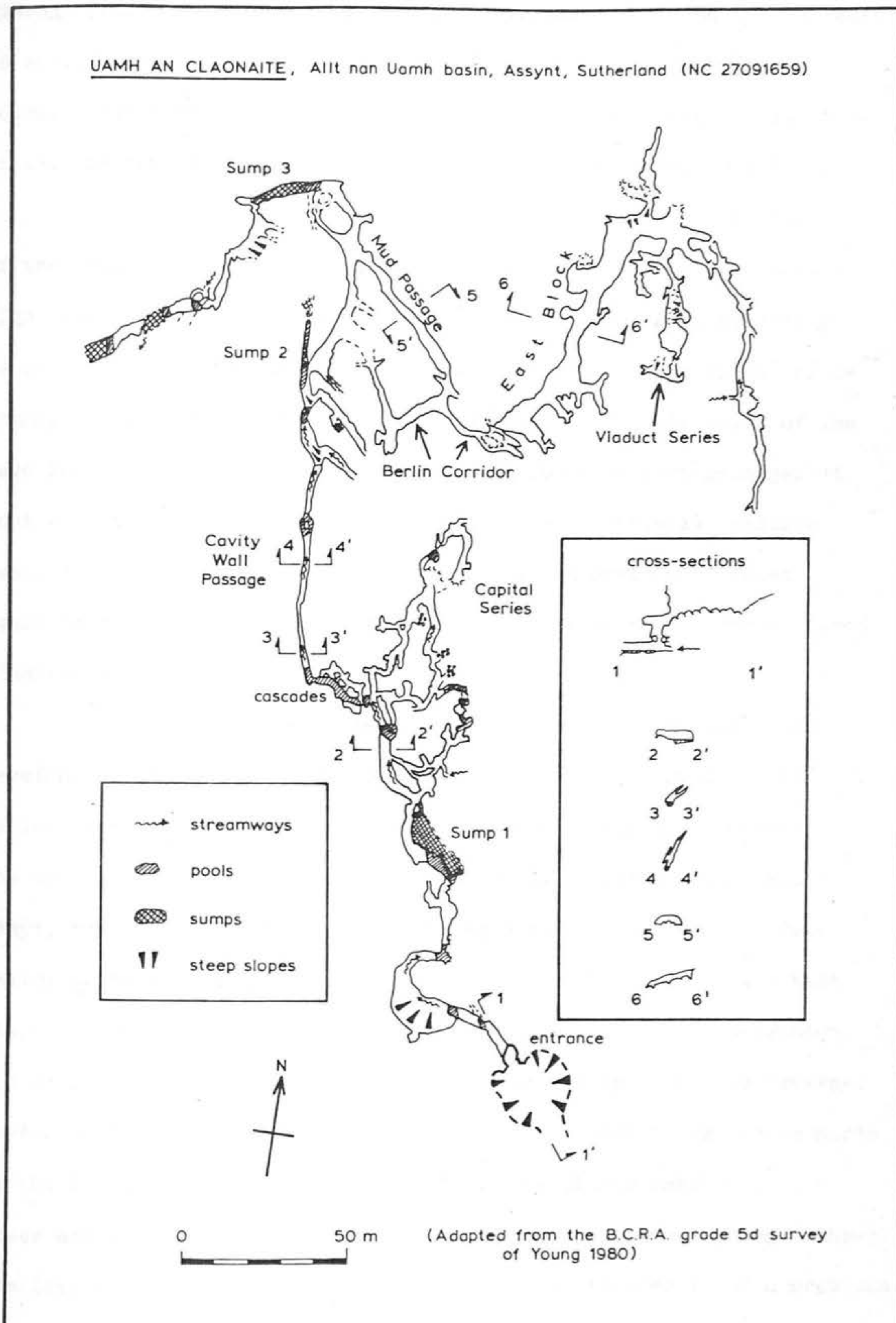
leading directly from Breakdown Chamber.

Current views on the development of Allt nan Uamh Stream Cave (Campbell 1977) favour a preglacial phreatic phase with water in the phreas entering at Breakdown Cavern, flowing down Oxford Street, and along and up to Drip Chamber; continuation of this route is presently blocked by a collapsed roof. Campbell believed that water would also have entered the cave at the Assembly Hall at this time, flowing down Piccadilly to join the main water flow, but this appears improbable under phreatic conditions; water flowing from Piccadilly to the Assembly Hall would seem more likely. A change to vadose conditions with the lowering of the local water-table, attributed by Campbell to glaciation, resulted in incision of new passages and the exploitation of joints and bedding planes to form a lower series of vadose passageways.

(ii) Uamh an Claonaite (NC 27091659). Scotland's largest cave to date, Uamh an Claonaite has been recently described by Jeffries & Young (1980) and Young (1980) (Fig. 9.10). Most of the passages can be attributed to a phreatic origin, though some have been drastically modified by breakdown and vadose incision. Most of the passageways appear to have developed along the strike of the beds, which in this area dip towards the east and NE. The influence of thrust and other fault planes on the configuration of the cave is difficult to determine as exposure of solid rock at the surface is restricted by extensive peat cover, resulting in little detail on the local geological map. Nevertheless, some structural control on several of the passageways is clearly evident from within the cave. A vosgesite dyke controls the orientation of the first section of streamway one negotiates on entering the cave. The inclined rift of Cavity Wall

Fig. 9.10 Survey of Uamh an Claonaite.

(diagram overleaf)



UAMH AN CLAONAITE, Allt nan Uamh basin, Assynt, Sutherland (NC 27091659)
Sump 3
Mud Passage
5
6
East Block
6'
5'
Sump 2
Viaduct Series
Berlin Corridor
Cavity Wall Passage
4
4'
Capital Series
cross-sections
3
3'
cascades
1
1'
2
2'
streamways
pools
sumps
steep slopes
Sump 1
3
3'
4
4'
5
5'
6
6'
1
entrance
N
1'
0
50 m
(Adapted from the B.C.R.A. grade 5d survey of Young 1980)

Passage probably follows a fault plane (although the throw of the fault is small), and it has been suggested (I.R. Young and others, pers. comms.) that Sump 3 is aligned along a fault which is the continuation of the one running behind the nearby Creag nan Uamh (Fig. 9.11).

The present active streamways occur in the lowest levels of the cave, progressively descending by way of a number of cascades. High-level phreatic passageways exist on two main levels, generally descending to the NW, as shown by initial results (unpublished) of an attempt by I.R. Young and the author to construct an elevation of the cave from Young's survey measurements. These high-level passages at most take trickles of water and are therefore essentially inactive features (although replacement of disturbed sediments with fresh deposits suggests that parts of the Capital Series take water in flood conditions).

Clastic deposits and speleothems indicate a multiphase development of the cave. Active streamways contain normal fluvial facies, with gravel deposits showing a high allochthonous content. The once phreatic Capital Series contains mainly sands, silts and clays, the finer sediments usually being a deep brown colour. Pale yellow silts and sands occur extensively in Mud Passage and the East Block, often overlying breakdown deposits. Large areas of breakdown are present, especially in the chamber between Sump 2 and Mud Passage, in Berlin Corridor, in the northern half of the East Block and in parts of the Viaduct Series (Fig. 9.10). In the breakdown chamber at the lower end of Mud Passage, large blocks of thick flowstone occur amongst the fallen breakdown blocks. Some are cemented to breccia of a previous breakdown phase and show a number of distinct periods of flowstone deposition separated by hiatuses. About 35 m up Mud Passage, a grotto contains eroded flowstone <u>in situ</u> on the bedrock floor with two thin layers of white calcite separated by a thin mud hiatus. This is overlain by breakdown, in turn covered by white flowstone which is itself broken into blocks. Two flowstone deposits are also visible on the walls in this area, one beneath and one above the sand and silt fill of the passage. Speleothems overlying fine sediments also exist at the southern end of the East Block.

Young (1980, p. 52) hypothesised a preglacial phreatic origin along joints, faults or thrust planes. He argued that the phreas was dipping steeply, although there is no evidence that such a situation existed. During the phreatic phase many of the passages received a sedimentary fill. Lowering of the water-table by meltwater erosion in the Allt nan Uamh valley was seen as initiating calcite deposition, breakdown of sections of cave passage and progressive vadose enlargement of deeper routes. Young suggested that at least two cycles of the above phases have occurred in the cave.

(iii) The Creag nan Uamh (approx. NC 268170). This crag of Eilean Dubh dolomite possesses three caves collectively known as the 'Bone Caves', and a number of niches and rockshelters. The Bone Caves are considered in greater detail in chapter 12; only the cave descriptions in relation to the Allt nan Uamh valley and its hydrology will be given in this chapter. Situated 45 m above the valley floor at the foot of the crag, the three Bone Caves (Badger Cave, Reindeer Cave and Bone Cave, from west to east) have wide, semicircular entrances and large entrance chambers. Solution pockets and tubes in the roofs of these entrance chambers testify to their phreatic origin. Two of the caves (Badger Cave and Reindeer Cave) have inner chambers, largely filled with pale yellow silts and sands, topped by breakdown deposits. Bone Cave possibly has an inner chamber, but existing deposits in the outer cave prevent access. A low adjoining opening (Foxes Den) is also choked with clastic deposits which may conceal another large cave behind.

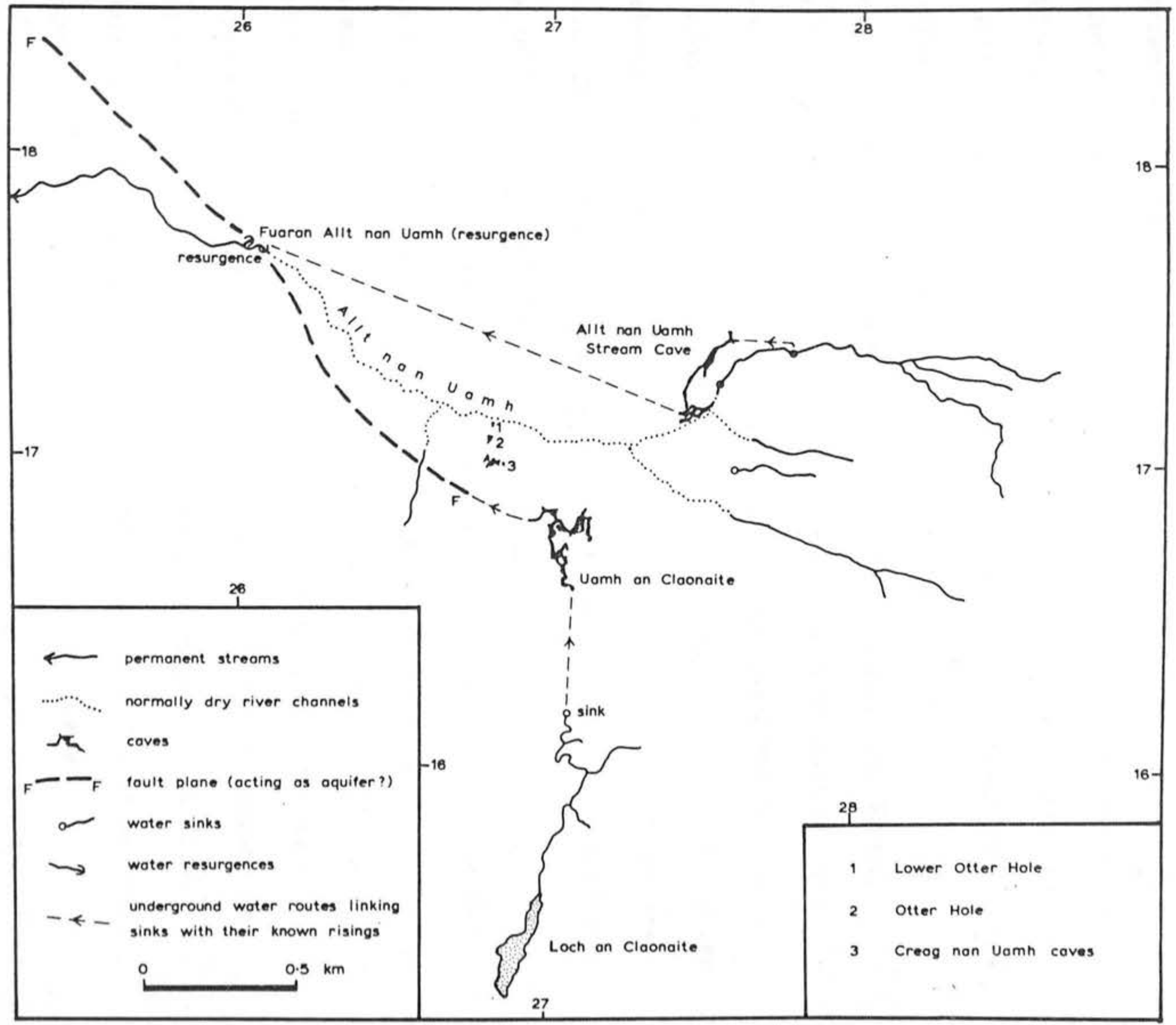
Peach & Horne (1917) saw the caves as resurgences at the time the Allt nan Uamh Valley was filled to that level with glacial deposits. They stated that at this point in the valley "there is a considerable development of drift ... On the northern slope it forms a well-marked terrace ... whose surface is about the level of the bone-cave. A corresponding terrace is observable about 1200 yards farther down the valley ... Water entering the limestone was obliged to escape at the edge of the terrace where it bounded the limestone crag" (Peach & Horne 1917, pp. 337-338). Water was later to resurge at Otter Hole, below the crag, when this drift terrace had been partially removed. However, there is no evidence of a terrace of drift at or about the level of the Bone Caves; glacial deposits in the main valley only occur at least 50 m below this level. It seems more reasonable to suggest that the Creag nan Uamh caves are the truncated remains of large, high-level phreatic passages forming part of a system that once stretched across the area now occupied by the Allt nan Uamh valley. It is interesting to note that the Bone Caves are at approximately the same altitude (330 m O.D.) as some of the large, abandoned phreatic passages in Allt nan Uamh Stream Cave and Uamh an Claonaite.

(b) Surface features and hydrology

The Allt nan Uamh drainage begins on the slopes of the southern part of the Breabag ridge and Creag Liath (Fig. 9.11). The Allt nan Uamh is a permanent stream whilst it flows over the Cambrian quartzites, but it sinks at various places during its course over the

Fig. 9.11 Hydrology of the Allt nan Uamh drainage basin.

(diagram overleaf)


Fuaran Allt nan Uamh (resurgence)
resurgence
Allt nan Uamh
Allt nan Uamh Stream Cave
Uamh an Claonaite
sink
Loch an Claonaite
F
26
27
28
18
17
16
permanent streams
normally dry river channels
caves
fault plane (acting as aquifer?)
water sinks
water resurgences
underground water routes linking sinks with their known risings
0
0·5 km
1 Lower Otter Hole
2 Otter Hole
3 Creag nan Uamh caves

dolomite and there is rarely more than a steady trickle of water falling over the waterfall at NC 27531722 (formed by the presence of an igneous intrusion). This water enters Allt nan Uamh Stream Cave where it forms the streams in the Farr Series and in the Rift Series of passageways. Two small south-bank tributary streams are usually dry for the lower part of their courses. Another small stream sinks at NC 27591699 in a boulder-choked swallet. The stream channel of the Allt nan Uamh, which is normally dry for 2 km downstream from the waterfall except in high discharge conditions, cuts across an old cave passage that was originally part of the Allt nan Uamh Stream Cave system. The portion left on the north bank forms part of the present cave; the portion left on the south bank now forms Bear Cave (NC 27481712), an inclined passageway almost completely full of angular breakdown rubble in a silty-clay matrix. In this cave were found the remains of brown bears (Ursus arctos); one femur yielded a radiocarbon date of 2,673 $\pm$ 54 (BM-724) (Burleigh et al. 1976).

A dry valley extends towards Loch an Claonaite along the east side of the Creag nan Uamh. This valley is entirely inactive, drainage waters from the loch sinking at NC 27061617, eventually to flow through Uamh an Claonaite. The peaty area between the loch and the present entrance to Uamh an Claonaite is pitted with shakeholes, some interconnected with small grassy-floored channels. In flood conditions several small resurgences and sinks are probably active within these otherwise dry networks of shallow depressions.

A dry channel leading from Otter Hole and Lower Otter Hole beneath the Creag nan Uamh indicates that these small caves once acted as resurgences, although only the latter cave is still operational in flood conditions. Several small resurgences occur from beneath the drift in the valley bottom at a number of places between the Creag nan

Uamh and the two main risings, Fuaran Allt nan Uamh (NC 26051773) and a small south-bank resurgence 100 m upstream. All drainage waters from the An Claonaite and upper basin areas resurge here, the Allt nan Uamh being a permanent river downstream of these points. These risings occur where the continuation of the major fault that runs behind the Creag nan Uamh crosses the main valley (Fig. 9.11), and where an impermeable vosgesite intrusion crops out. 0.5 km downstream of the risings, waterfalls have been produced as the river crosses the vosgesite.

9.2.3 The Abhainn a' Chnocain basin

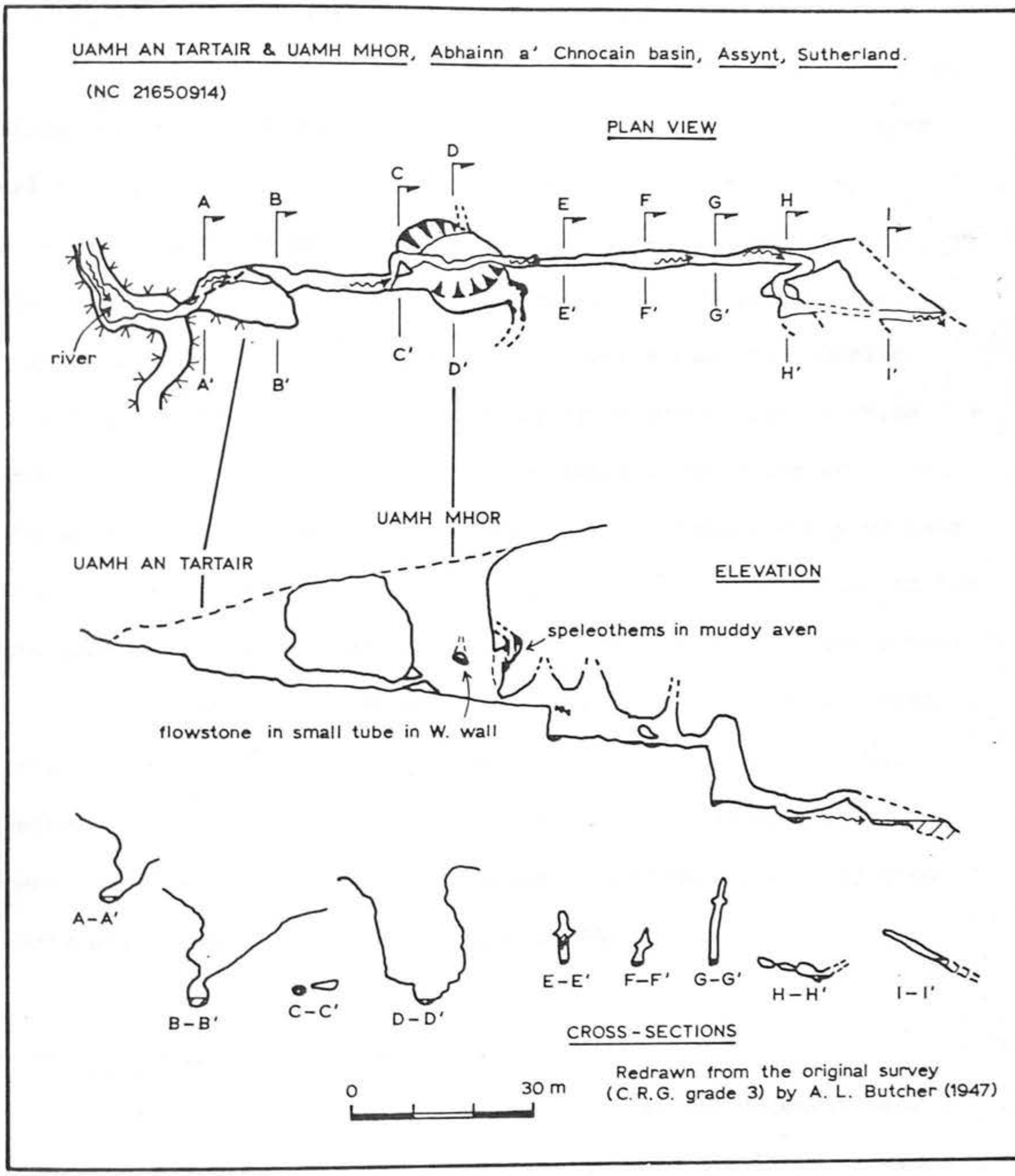
(a) Major caves

Jeffries (1972) and Dowswell (1976) gave brief descriptions of the major caves in this basin. The larger caves have all developed along rifts associated with lines of weakness in the bedrock. None of the caves have much horizontal development, although several are quite deep (*e.g.* Uamh an Poll Eoghainn (NC 206093) has a depth of 23.6 m and Elphin Hole (NC 210095) is nearly 20 m deep).

The one cave in which the writer found a reasonable lithostratigraphy, including dateable speleothems, was Uamh an Tartair (Knockan) (not to be confused with Uamh an Tartair (Traligill)). This cave has developed along a fault plane inclined at about 35° to the west (Fig. 9.12). It includes the large pothole of Uamh Mhor, *c*. 20 m to the north of the entrance, which is the result of de-roofing of part of the cave after the removal of quartzite that capped the dolomite in this area. A small solutional tube on the western wall of Uamh Mhor and a large aven on the eastern side have phreatic affinities, although breakdown and vadose incision have greatly affected the

Fig. 9.12 Survey of Uamh an Tartair and Uamh Mhor in the Abhainn a' Chnocain drainage basin.

(diagram overleaf)


UAMH AN TARTAIR & UAMH MHOR, Abhainn a' Chnocain basin, Assynt, Sutherland.
(NC 21650914)
PLAN VIEW
A B C D E F G H I
A' B' C' D' E' F' G' H' I'
river
UAMH MHOR
UAMH AN TARTAIR
ELEVATION
speleothems in muddy aven
flowstone in small tube in W. wall
A-A'
B-B'
C-C'
D-D'
E-E'
F-F'
G-G'
H-H'
I-I'
CROSS-SECTIONS
0
30 m
Redrawn from the original survey
(C.R.G. grade 3) by A. L. Butcher (1947)

appearance of much of the remainder of the cave. The stream that enters the mouth of the cave and crosses the floor of Uamh Mhor enters a narrow rift (aligned along the fault which is nearly vertical here), descending to a sump by way of a series of cascades.

In the phreatic pocket on the west side of Uamh Mhor, 2 cm of white crystalline flowstone are separated from another 1 cm layer of calcite by an erosion surface and 1-2 mm of silty mud. This in turn has been facetted by resolution, probably representing the filling of the cave with water to at least this level. 15 cm of a quartzite-rich gravel, cemented with a buff-coloured impure calcite, overlie this layer, in turn capped with a 5 mm layer of dirty calcite which is undergoing resolution at present. The dense crystalline structure of the white flowstone layers at the base of this sequence may predate the unroofing of Uamh Mhor, for speleothems presently deposited in the photic zone are of more 'sugary' appearance (T.C. Atkinson, pers. comm.).

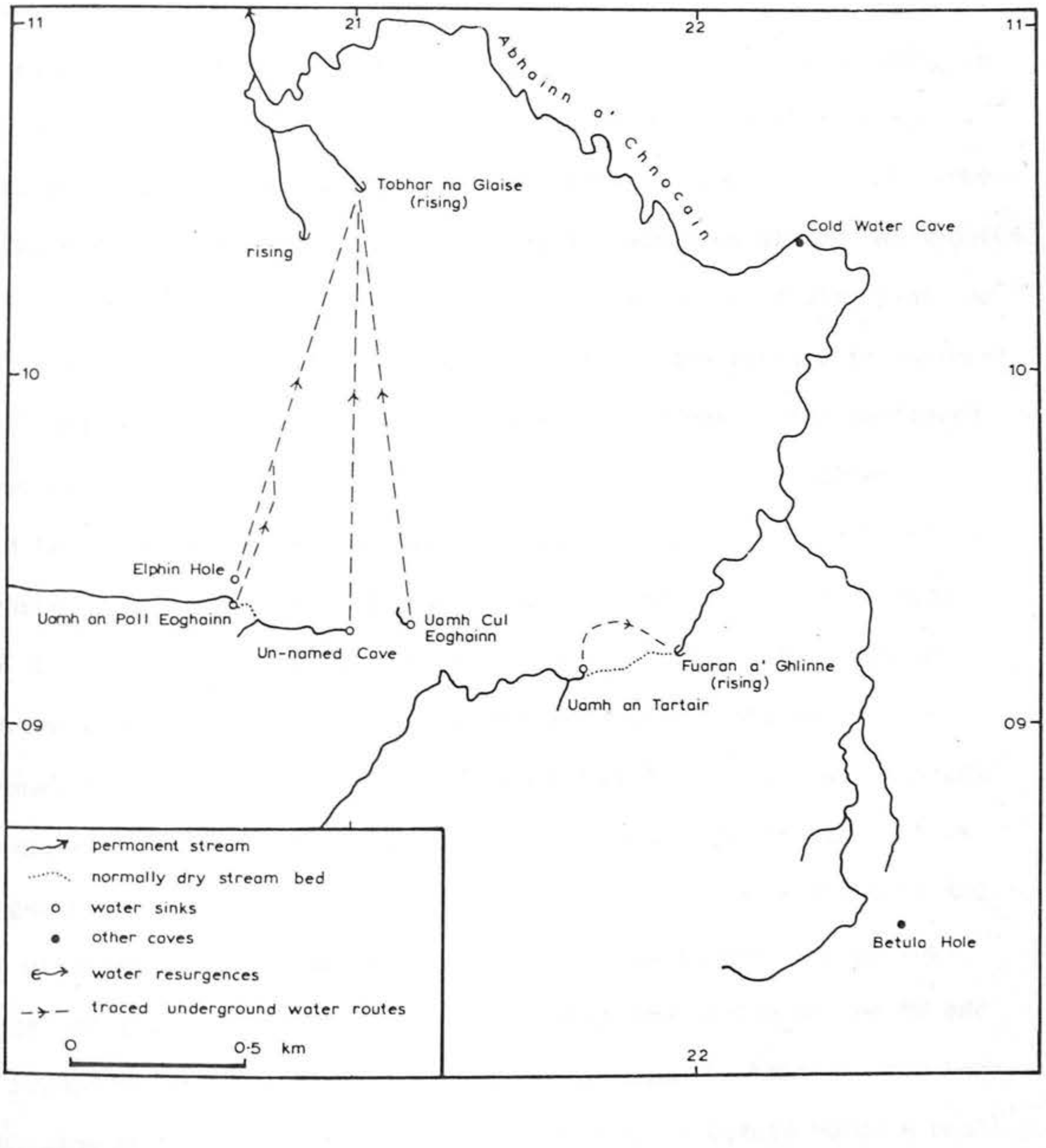
In the muddy aven on the east side of Uamh Mhor a grotto contains several large columns and some stalagmites which have been deposited on breakdown blocks and subsequently covered with silts. A further flowstone phase covers these fine sediments, in turn covered by peaty silts and flood debris of the present day.

(b) Surface features and hydrology

The hydrology of the area (Fig. 9.13) has recently been described by Dowswell (1976). The Abhainn a' Chnocain is initiated near Loch Odhar from where it flows to the NNE over the Moine schists. On reaching the Moine Thrust, it flows towards the east over quartzite before turning abruptly northwards to enter Uamh an Tartair where a fault occurs in the dolomite. The dry river bed, occupied in high discharge conditions, continues over a waterfall until NC 21950918

Fig. 9.13 Hydrology of the Abhainn a' Chnocain drainage basin.

(diagram overleaf)


Abhainn a' Chnocain
Tobhar na Glaise
(rising)
rising
Cold Water Cave
Elphin Hole
Uamh an Poll Eoghainn
Un-named Cave
Uamh Cul
Eoghainn
Fuaran a' Ghlinne
(rising)
Uamh an Tartair
Betula Hole
permanent stream
normally dry stream bed
water sinks
other caves
water resurgences
traced underground water routes
0
0·5 km
11
21
22
10
09

where the waters return from Uamh an Tartair at the Fuaran a' Ghlinne rising. From this point, the Abhainn a' Chnocain flows as a permanent stream until it reaches the eastern end of Loch Veyatie. A small oxbow cave, Cold Water Cave, cuts across a bend in the river at NC 22281038.

A tributary stream starts to the east of Cnoc an t-Sasunnaich and runs along the outcrop of the Moine Thrust plane before sinking in Uamh an Poll Eoghainn (NC 20660933). This water, plus that from other sinks such as Un-Named Cave (approx. NC 210093), Elphin Hole (NC 210095) and Uamh Cul Eoghainn (NC 205094), rises at Tobhar na Glaise (NC 21031053).

With a large proportion of this area covered with peat, up to 3 m deep in places, surficial karst features are limited to numerous small dolines and a few instances of karren features on the scattered areas of bare dolomite bedrock. In no instances can the latter be said to constitute limestone pavements as they are very restricted in occurrence and area. One notable geomorphic feature is a 10 m-deep dry channel, Feadan Mor, at approximately NC 227082. It is cut in bedrock with dolomite to the north and quartzite to the south. The channel cuts across the ridge which separates the Abhainn a' Chnocain drainage basin from the streams that drain into Loch Urigill. It is suggested that Feadan Mor was initially cut by glacial meltwaters and was utilised by proglacial streams when ice lay to the east of the ridge through which it is cut. Two 2 m-deep shakeholes at the NW end of Feadan Mor suggest that water flowing through the channel sank here. A shallow channel continues to the NW, leading to Betula Hole, a small cave at NC 22600836.

9.3 SYNTHESIS

The survey of the various cave systems in the area has shown that most, if not all, of the caves were initially formed phreatically when local water-tables and, presumably, the local land surface were much higher than at present. Subsequent roof collapse and vadose incision have drastically altered the appearance of the caves. All the more extensive cave systems possess large, high-level phreatic passages which are in several places choked with a sedimentary fill to their roofs. Active streamways tend to be confined to lower levels. The configuration of most of the caves owes much to the various lines of weakness in the local bedrock, particularly to faults and thrust planes formed by the intense tectonic movements that have affected the region. In the Traligill basin, one thrust plane - the Traligill Main Thrust - has played a large part in the formation of the present drainage system.

Cave sediments in the area can be sub-divided into three main categories : breakdown deposits, formed by cavern collapse; rounded cobbles, gravels and sands comprising fluvial facies; and fine-grained deposits, some laminated, indicating deposition in quiet-water conditions. The latter category can be further sub-divided into the pale yellow silts and sands that occur in the large, abandoned phreatic passageways, and other fine deposits which tend to be darker in colour and present at lower levels in the cave.

Speleothems are present above and below certain of the lithostratigraphic units, and incorporated in others. Their presence therefore indicates that several discrete depositional episodes have occurred during the development of the subterranean drainage network during which conditions were relatively stable for long periods of time.

The mapping of surficial geomorphic features in the various karst drainage basins has shown that a progressive lowering of the water-table has resulted in suites of dry channels (some permanently dry, others only seasonally dry) and a number of water resurgences and sinks.

Further work was therefore undertaken to attempt to resolve the following problems:

(a) the age of various parts of the cave systems;

(b) the reasons for, and timing of, the progressive lowering of the local water-table;

(c) the provenance and mode of deposition of the various sedimentary groups, and their bearing on the environmental history of the area.

This work is discussed in chapters 10 and 11.

CHAPTER 10 : ANALYSIS OF THE CAVE SEDIMENTS

10.1 INTRODUCTION

A prerequisite to understanding the provenance and mode of deposition of the different sedimentary units in the caves was an analysis of various sedimentary properties. Several representative bulk samples were taken from the caves and surrounding areas. These are listed in Appendix II. Particle size analysis was undertaken for all the sediment samples, and samples from the various gravel layers and tills were analysed for lithological composition and roundness.

10.2 ANALYTICAL TECHNIQUES

10.2.1 Particle Size Analysis

Admixtures of sticky, clay-rich deposits and coarse particles are common in the cave environment. In such cases normal dry-sieving techniques are unsatisfactory. The method adopted for the analysis of samples from the Assynt caves followed fairly closely the wet-sieve replicate method of Folk (1974) and similar techniques advocated by Buller & McManus (1979).

All field samples were air-dried in the laboratory, then disaggregated in a mortar, using a rubber pestle to minimise damage to clay-sized particles. Disaggregation continued until the total sample could be passed through a sieve with a mesh of 2 mm (-1ϕ), excluding rock particles with greater diameters. These rock particles greater than -1ϕ were sieved through a stack of sieves with meshes at whole-phi intervals, and the weights retained on each sieve were expressed as percentages of the total sample. The remainder of the disaggregated sample was split using the coning-and-quartering technique until a representative sub-sample of 80-150 g was obtained, dependent on the coarseness of the sample.

Each sub-sample was accurately weighed to 0.1 g on a torsion balance, then dispersed overnight in a solution of 33 g of sodium hexametaphosphate and 7 g of sodium carbonate, made up to 1000 ml with distilled water. After vigorous stirring, the sample and dispersant were washed through a 63 μm (4ϕ) mesh until all the silt- and clay-sized particles were washed out. This material passing through the sieve (hereafter referred to as 'the fine fraction') was collected and placed in a 1000 ml measuring cylinder for further analysis. The fraction retained on the sieve (hereafter referred to as 'the coarse fraction') was dried in an oven at about 110°C for 24 hours and then allowed to cool in a dessicator to prohibit the uptake of hygroscopic moisture.

The oven-dried coarse fraction had to be related to the weight of the original sample, which was air-dried; hence further analysis of this fraction was preceded by the correction of the original sub-sample weight to one that made allowance for hygroscopic moisture. About 100 g of the field sample were weighed to 0.1 g, dried in an oven at 110°C for 24 hours and cooled in a dessicator prior to reweighing. The loss of moisture on heating was expressed as a percentage of the original weight; this percentage was then deducted from the original air-dried weight of the sub-sample to give a corrected 'oven-dried' weight. The coarse fraction of the sub-sample was dry-sieved through a nest of sieves at half-phi intervals. Any material passing through the bottom sieve (4ø) was added to that already in the measuring cylinder. The material retained on each sieve was measured to 0.1 g and expressed as a percentage of the corrected sub-sample weight. Total weights of the coarse and fine fractions were calculated and expressed as percentages of the corrected sub-sample weight. The fine fraction was analysed using the pipette method (e.g. Folk 1974; Buller & McManus 1979), whereby aliquots of a suspension of the silts and clays at 20°C were taken from set depths in the liquid at specific times, and oven-dried weights obtained for the contained sediments. Use was made of a sampling pipette attached to a fixed stand to minimise disturbance of particles on insertion into the suspension, as recommended in British Standard 1377 (1975).

A total of 59 samples was analysed by this method.

10.2.2 Lithological Analysis

The -2ø to -3ø size fraction was analysed for lithological composition after being separated from the rest of the sample during the particle size analysis. From preliminary studies, four main classes could be distinguished on lithological grounds : (i) dolomite, (ii) quartzite and Torridonian, (iii) vosgesite and (iv) a miscellaneous group containing all the other lithologies. Where possible, a minimum of 300 particles was counted for each sample. A total of 16,034 stones from 32 samples was counted during the course of this work.

12 samples of the fine sediments were analysed by X-ray diffraction to determine their mineralogy (analysis kindly undertaken by P. Aspen in the Grant Institute of Geology, Edinburgh, and results supplied by S.A. Caswell of the Geology Department, Sheffield University).

The same samples were also analysed by acetolysis in a 10% solution of dilute acetic acid; subsequent weight loss was taken to represent their carbonate content, and hence the proportion of the sample composed of $CaCO_3$.

10.2.3 Roundness Analysis

Roundness of a particle is dependent on mode and distance of transportation, its lithology and its size (e.g. Krumbein 1941; Pettijohn 1957). In view of these last two factors, dolomite and quartzite particles from the -2ø to -3ø size fraction of each of the cave gravel and till samples were analysed separately for roundness. For speed of analysis, use was made of the Powers' visual comparison charts (Powers 1953) which show six classes of increasing roundness at two levels of sphericity. A total of 6,000 stones was analysed in this way. Following the rho-transformation method of Folk (1955), inclusive graphic statistical techniques enabled the calculation of mean roundness ($\bar{x}_\rho$) and standard deviation (σ_ρ) for each sample to allow results to be compared on a scale from zero, indicating very angular particles, through to 6.0, indicating well-rounded particles.

10.3 RESULTS

10.3.1 Fine deposits

In the Assynt caves the sub-division of fine sediments into modern flood deposits and fossilised fine deposits is based on both geomorphic evidence and colour of the sediments. Flood deposits are not found in the high-level phreatic passages of the larger cave systems, being restricted to those parts of a cave above the level of the active streamways where quiet-water conditions exist when backing-up of water occurs on flooding. They are commonly medium brown or deep brown in colour, often laminated and possessing sedimentary structures indicative of a unidirectional water flow (e.g. ripples and smallscale cross-bedding). The fossilised fine deposits are paler in colour (Munsell colours of 10YR 5/4 and 10YR 6/3 being most often recorded) and occupy many of the large high-level passages that are not prone to modern flooding. In some cases, these large passages are filled almost to the roof with these deposits (e.g. Reindeer Cave on the Creag nan Uamh, the East Block in Uamh an Claonaite and Rabbit Warren in the Cnoc nan Uamh cave system) and in other locations there is evidence that passageways formerly contained much more of these deposits than they do at present (e.g. Oxford Street in Allt nan Uamh Stream Cave). Only the sediments studied in the section near to the Stream Chamber in Uamh an Tartair (Traligill) were clearly laminated; all other examined sections lacked signs of bedding. This group of sediments tended to be relatively dry and friable compared to many of the water-charged, sticky muds found in the cave environment. A progressive drying-out of these sediments may have removed any visual contrast between adjoining laminae, especially if differences in grain size and shape were minimal.

Particle size analysis of the fossilised fine deposits has shown that, of 17 bulk samples, 11 are unimodal with median particle sizes ranging from 4.9ϕ to 7.9ϕ (i.e. silt-sized particles). Three samples taken from the inner Reindeer Cave were multimodal and three from Badger Cave were bimodal. (These sediments from the Creag nan Uamh caves are considered in greater detail in chapter 12.) There is evidently a degree of variation between samples from different localities despite similarities in outward appearance.

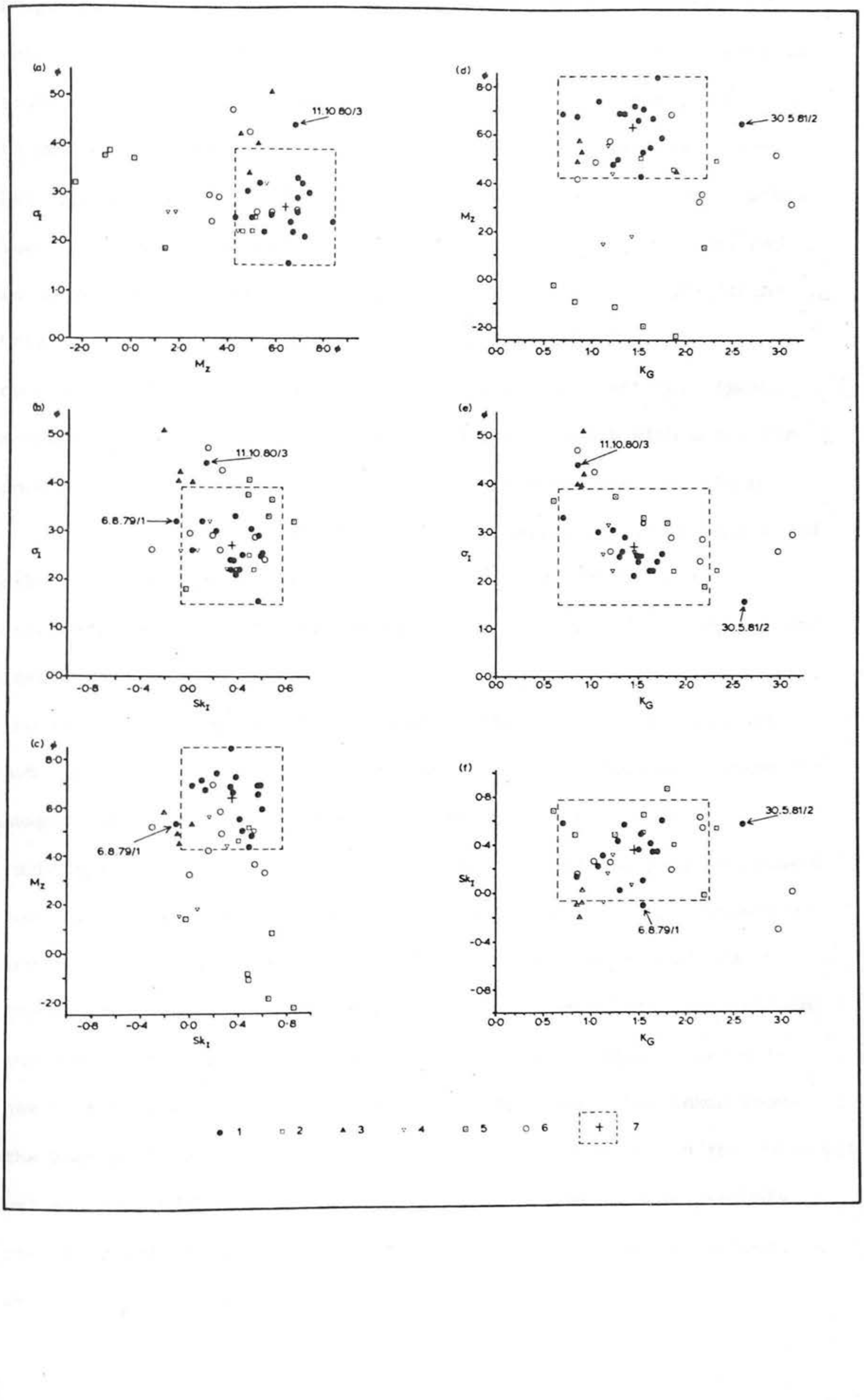
Size parameters of mean (M_Z), sorting (σ_I), skewness (SK_I) and kurtosis (K_G) were calculated from cumulative percentage frequency curves drawn from the particle size data for each sample, using the inclusive graphic statistical technique of Folk & Ward (1957) with McCammon's (1962) calculation for the mean (considered to be more representative of the total sample (Folk 1966)). Bivariate scattergram plots of the different size parameters are presented in Fig. 10.1. A zone was drawn two standard deviations (2σ) around the mean value of each of the size parameters of the 17 samples of the fossil fine deposits to determine the degree of similarity of the different samples. Most of the samples plotted within the 2σ-zone, but three samples did not : 11.10.80/3 (reddish-yellow silts, Reindeer Cave) was more poorly sorted than the rest ($\sigma_I = 4.36$); 6.8.79/1 (upper sample of yellow silts, Reindeer Cave) was slightly negatively skewed ($SK_I = -0.11$); 30.5.81/2 (yellow silts from the East Block, Uamh an Claonaite) was very leptokurtic ($K_G = 2.61$). The bivariate plots also show the size parameters of the samples of other sediment types for comparison with the fossil fine cave deposits. Many of the other sediment types show similarities, plotting within the 2σ-zone of the latter group, especially the modern cave flood deposits and some of the fluvial cave sands. Fine deposits taken from sections on the surface (e.g. river alluvium) plot variously around and within the 2σ-zone. The graph of σ_I against M_Z (Fig. 10.1(a)), used successfully by others to differentiate between different depositional environments (e.g. Friedman 1961), separates the coarse fluvial cave deposits and the silty clays interpreted as wash deposits in Bone Cave and Uamh Cailliche Peireag from the fossil cave silts (although one of the Bone Cave wash deposit samples falls just within the 2σ-zone).

The relative proportions of sand, silt and clay in each

Fig. 10.1 Bivariate scattergrams of various size parameters of certain sediment samples from the Assynt area.

Key:
1. Fossil cave silts.
2. Cave flood deposits.
3. Cave wash deposits.
4. Fine-grained cave stream deposits.
5. Coarse-grained cave stream deposits.
6. Fine-grained deposits from elsewhere in the study area (above ground).
7. Mean value for fossil cave silt samples, surrounded by a $\pm$ 2σ zone (95% confidence limits).

(diagram overleaf)


(a)
(b)
(c)
(d)
(e)
(f)
11.10.80/3
6.8.79/1
30.5.81/2
M_z
σ_I
Sk_I
K_G
ϕ
1
2
3
4
5
6
7

sample were calculated and the results presented as a ternary diagram (Fig. 10.2). A 2σ-zone was delimited around the plot of the mean values of the constituent size fractions of the fossil fine deposits (sand = 21.2%, silt = 55.2% and clay = 23.6%). Once more, a certain amount of inter-sample variation is apparent from the size of the 2σ-zone which occupies about one third of the total ternary diagram. Two samples plot outside the 2σ-zone : 4.7.79/2b (Badger Cave) had an above-average sand content, and 12.4.81/3 (Allt nan Uamh Stream Cave) had no sand-sized particles at all. Fig. 10.2 also shows the constituent size fractions of other sampled fine deposits. Again, similarities with the fossil cave silts are apparent with a certain amount of overlap occurring between the different sediment groups.

Ten samples of the fossil cave silts and seven samples of other types of fine sediment from the area were analysed for carbonate content. The results are shown in Table 10.1. Four of the fossil cave silt samples contained more than 10% carbonate by weight, the sample from the East Block, Uamh an Claonaite, containing over 30% (although a great deal of roof drip occurs at the point where the sample was collected, which may have artificially increased the carbonate content). In the other fossil cave silt samples, carbonate content is relatively low, the bulk of each sample being composed of insoluble particles. The two surface alluvium samples had low carbonate contents as might be expected from their locations near to the quartzite-dolomite boundary. The other analysed cave deposits had carbonate contents ranging from 13-21%. Grey silts taken from the base of karren features on the exposed thrust plane in the Traligill valley had a carbonate content of nearly 40%; these silts probably result largely from mechanical weathering of the dolomite surface, as well as a certain amount of solution.

Fig. 10.2 Ternary diagram of the percentage of sand, silt and clay in some fine-grained deposits from the Assynt area.

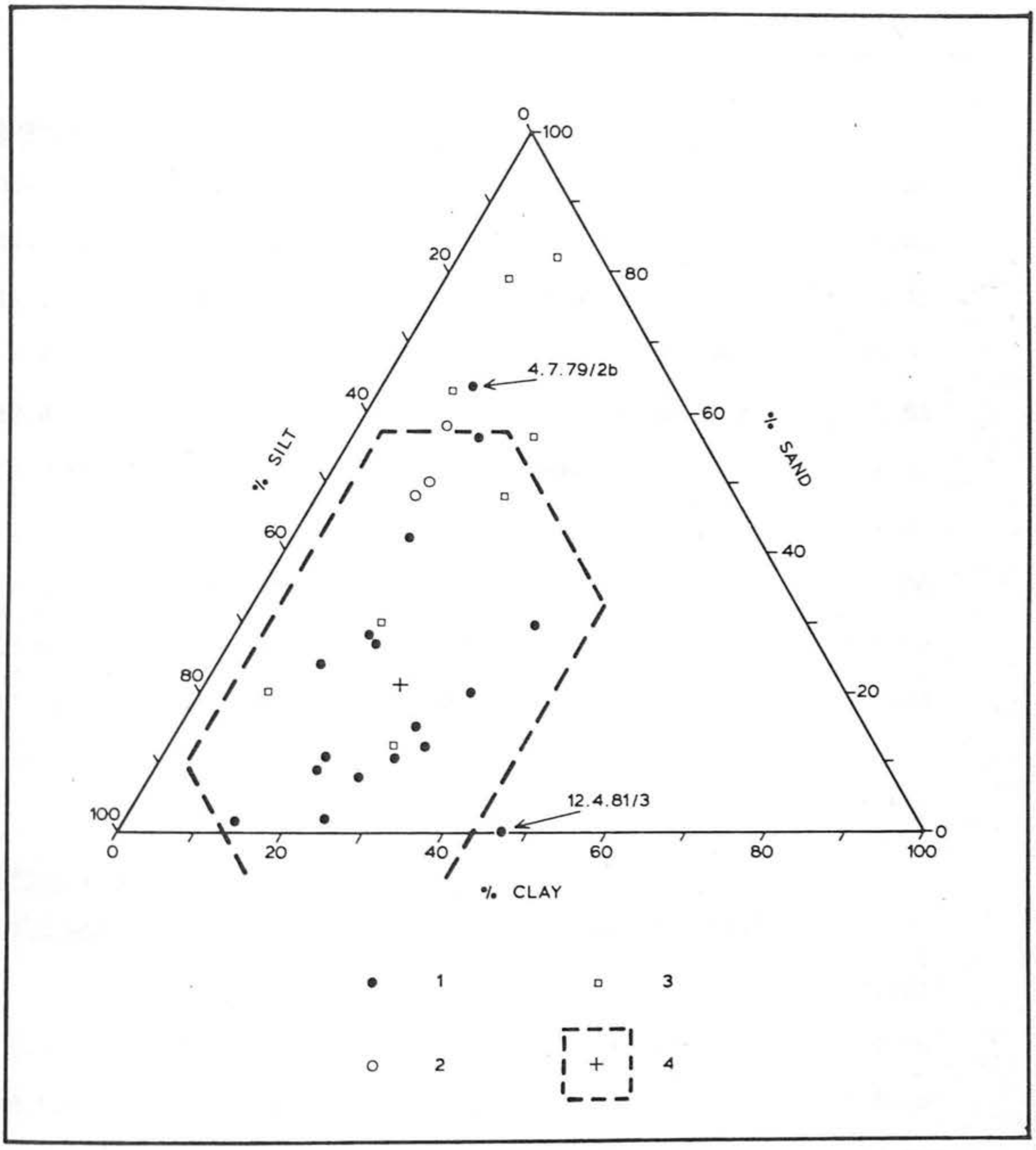
Key: 1. Fossil cave silts.

2. Cave flood deposits.

3. Fine-grained deposits from above ground.

4. Mean value for fossil cave silt samples, surrounded by a $\pm 2\sigma$ zone (95% confidence limits).

(diagram overleaf)



0
100
20
80
40
60
60
40
80
20
100
0
0
20
40
60
80
100
% SILT
% SAND
% CLAY
4.7.79/2b
12.4.81/3
1
2
3
4

TABLE 10.1. Carbonate content of fine sediment samples from the Assynt area.

	Sample	% Carbonate
Fossil cave silts		
30.5.81/2	East Block, U. an Claonaite	32.57
11.10.80/2	(Pale silts) Inner Reindeer Cave	18.62
12.10.80/3	Landslip Chamber, Cnoc nan Uamh	14.31
12.4.81/2	Oxford Street, Allt nan Uamh Stream Cave	10.91
12.4.81/3	Oxford Street, Allt nan Uamh Stream Cave	5.63
11.7.81/1	Viaduct Series, U. an Claonaite	5.32
4.7.79/4	Badger Cave	4.39
12.7.81/1	Uamh an Tartair (Traligill)	3.96
28.5.81/2	Uamh an Tartair (Traligill)	3.39
11.10.80/3	(Reddish silts) Inner Reindeer Cave	0.03
Other fine deposits		
29.3.80/1	Silts from exposed thrust plane, Traligill valley	39.63
29.5.81/1	Lower Traligill Cave, flood deposits	20.95
18.5.80/2	U. an Claonaite, flood deposits	15.43
28.5.81/5b	U. an Tartair (Traligill), red silts	14.83
31.5.80/1	Bone Cave, wash deposits	13.78
11.4.81/4	Claonaite flood channels, alluvium	3.55
14.8.80/1	Allt a' Bhealaich, alluvium	1.40

The results of the XRD analyses show that the mineralogical composition of the fossil cave silts is predominantly quartz with varying amounts of dolomite. Clay minerals (illite and chlorite) and feldspars are also present in very small quantities (S.A. Caswell, pers. comm.).

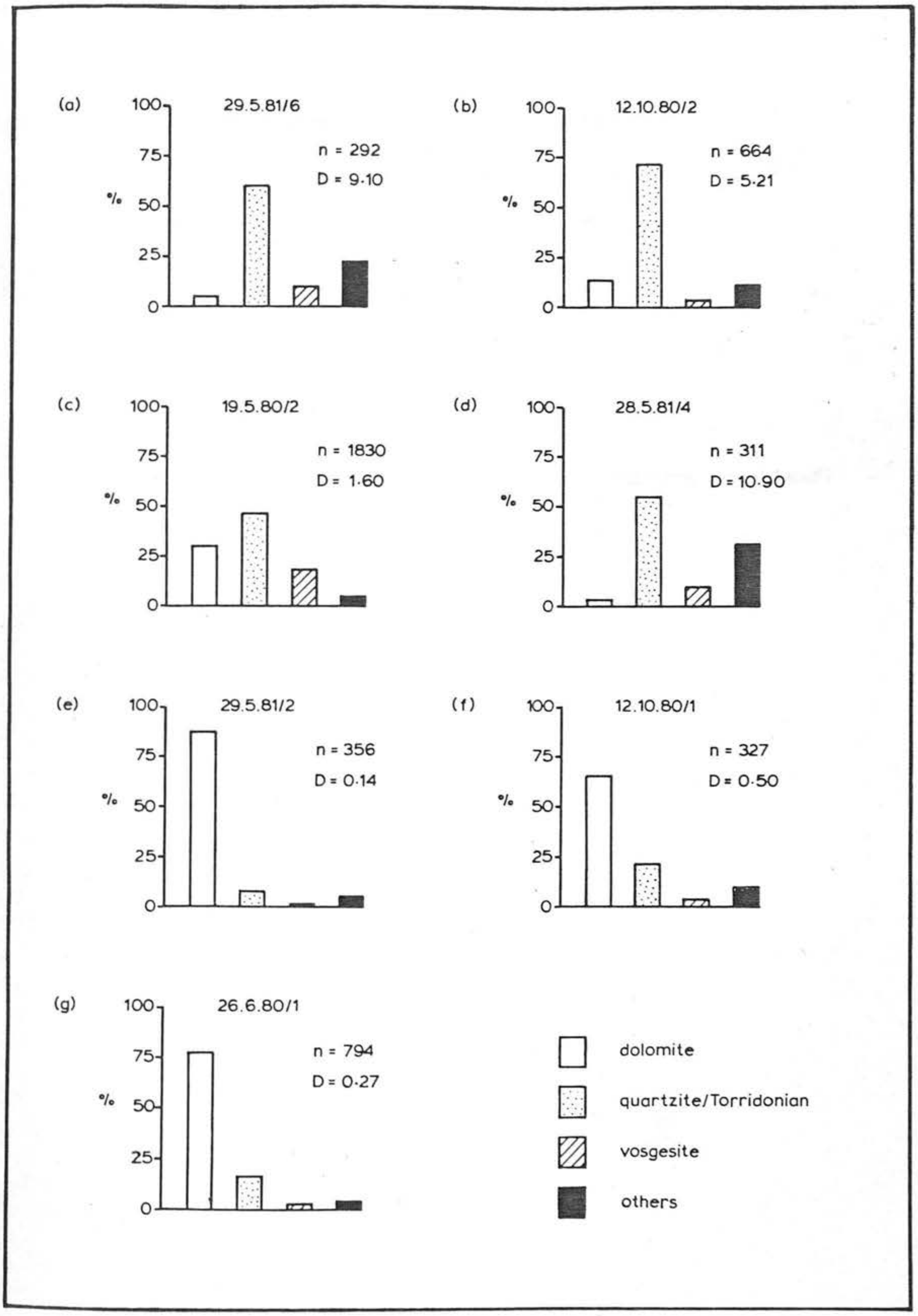
10.3.2 Gravels

Seven gravel samples from various caves, and six samples taken from till sections in the Traligill and Allt nan Uamh basins, will be considered here. Other gravel samples from the Creag nan Uamh caves will be discussed in chapter 12. The results of the lithological analyses on the 4-8 mm fraction of the cave gravels are shown in Fig. 10.3 and those for the till samples in Fig. 10.4. The cave gravels appear to fall into two groups : one shows a clear preponderance of quartzite/Torridonian stones with varying lesser amounts of other lithologies (Fig. 10.3(a), (b) and (d)); in the other dolomite clasts are dominant (Fig. 10.3(e), (f) and (g)), with the sample from Uamh an Claonaite (Fig. 10.3(c)) being somewhat transitional. It is suggested that the second group reflects the inclusion of a large amount of autochthonous breakdown deposits. This suggestion is supported by the results of the roundness analyses of these samples and the lithological analysis of the local till. Fig. 10.4 quite clearly shows that the till samples contained little or no dolomite clasts. The decalcified nature of the till was emphasised by the lack of any reaction when small 100 g sub-samples of the less than -1ϕ fraction were immersed in 10% dilute hydrochloric acid. It is therefore reasonable to assume that any dolomite stones in the Assynt cave gravels were internally derived. The roundness analysis of the dolomite-rich cave gravels (Fig. 10.3(c), (e), (f) and

Fig. 10.3 Results of the lithological analysis of the 4-8 mm size fraction of fluvial gravels from certain Assynt caves.

(<u>N.b.</u> for explanation of 'D'-values, see text.)

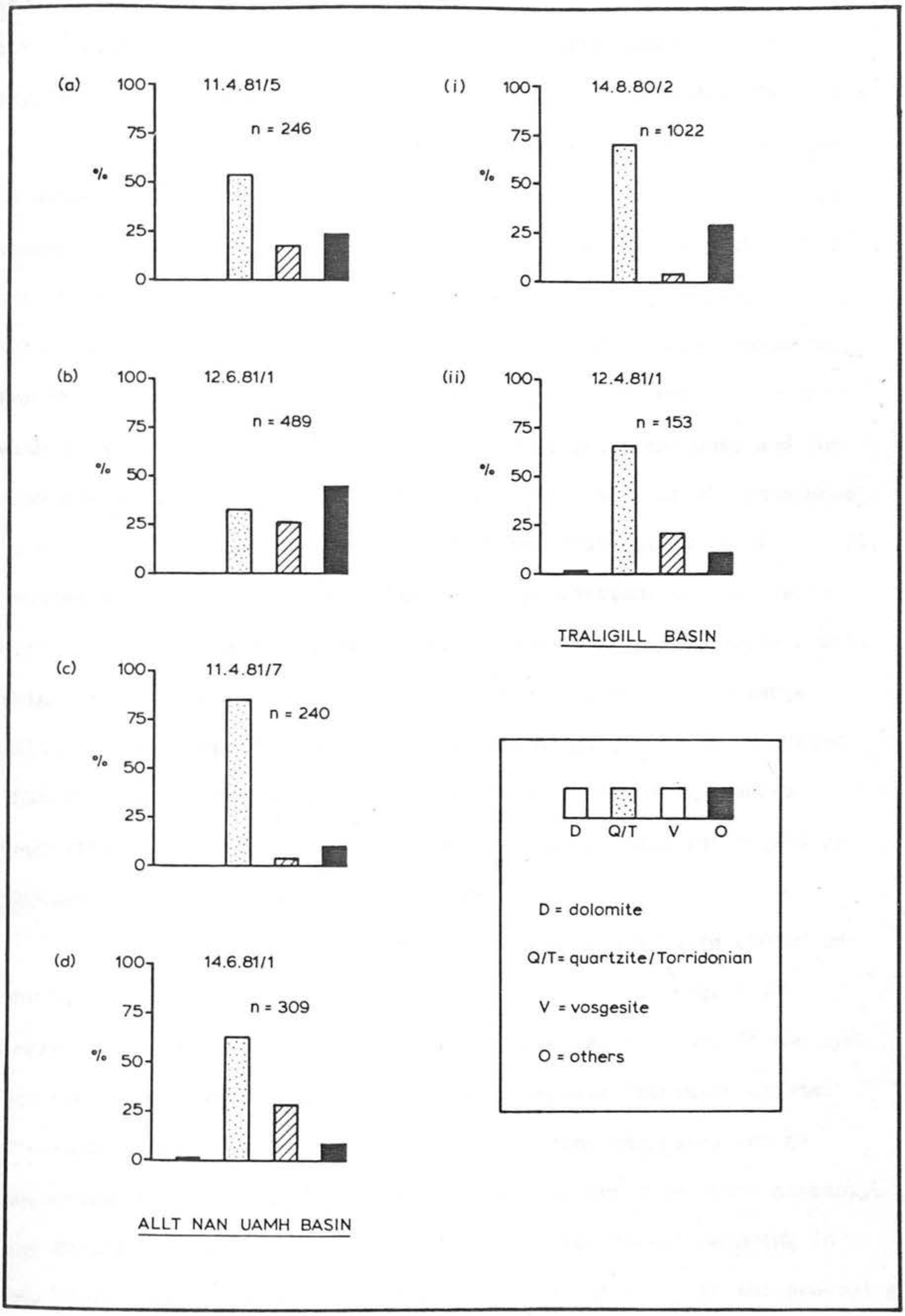
(diagram overleaf)



(a)
29.5.81/6
n = 292
D = 9.10
(b)
12.10.80/2
n = 664
D = 5.21
(c)
19.5.80/2
n = 1830
D = 1.60
(d)
28.5.81/4
n = 311
D = 10.90
(e)
29.5.81/2
n = 356
D = 0.14
(f)
12.10.80/1
n = 327
D = 0.50
(g)
26.6.80/1
n = 794
D = 0.27
%
0
25
50
75
100
dolomite
quartzite/Torridonian
vosgesite
others

Fig. 10.4 Results of the lithological analysis of the 4-8 mm size fraction of certain till samples from the Allt nan Uamh and Traligill drainage basins.

(diagram overleaf)


(a)
11.4.81/5
n = 246
(b)
12.6.81/1
n = 489
(c)
11.4.81/7
n = 240
(d)
14.6.81/1
n = 309
ALLT NAN UAMH BASIN
(i)
14.8.80/2
n = 1022
(ii)
12.4.81/1
n = 153
TRALIGILL BASIN
%
100
75
50
25
0
D
Q/T
V
O
D = dolomite
Q/T= quartzite/Torridonian
V = vosgesite
O = others

(g)) gave mean roundness figures of 1.81 $\pm$ 0.50, 1.69 $\pm$ 0.84, 1.97 $\pm$ 0.88 and 1.32 $\pm$ 0.31 respectively, illustrating the high angularity of these stones as would be expected of breakdown fragments.

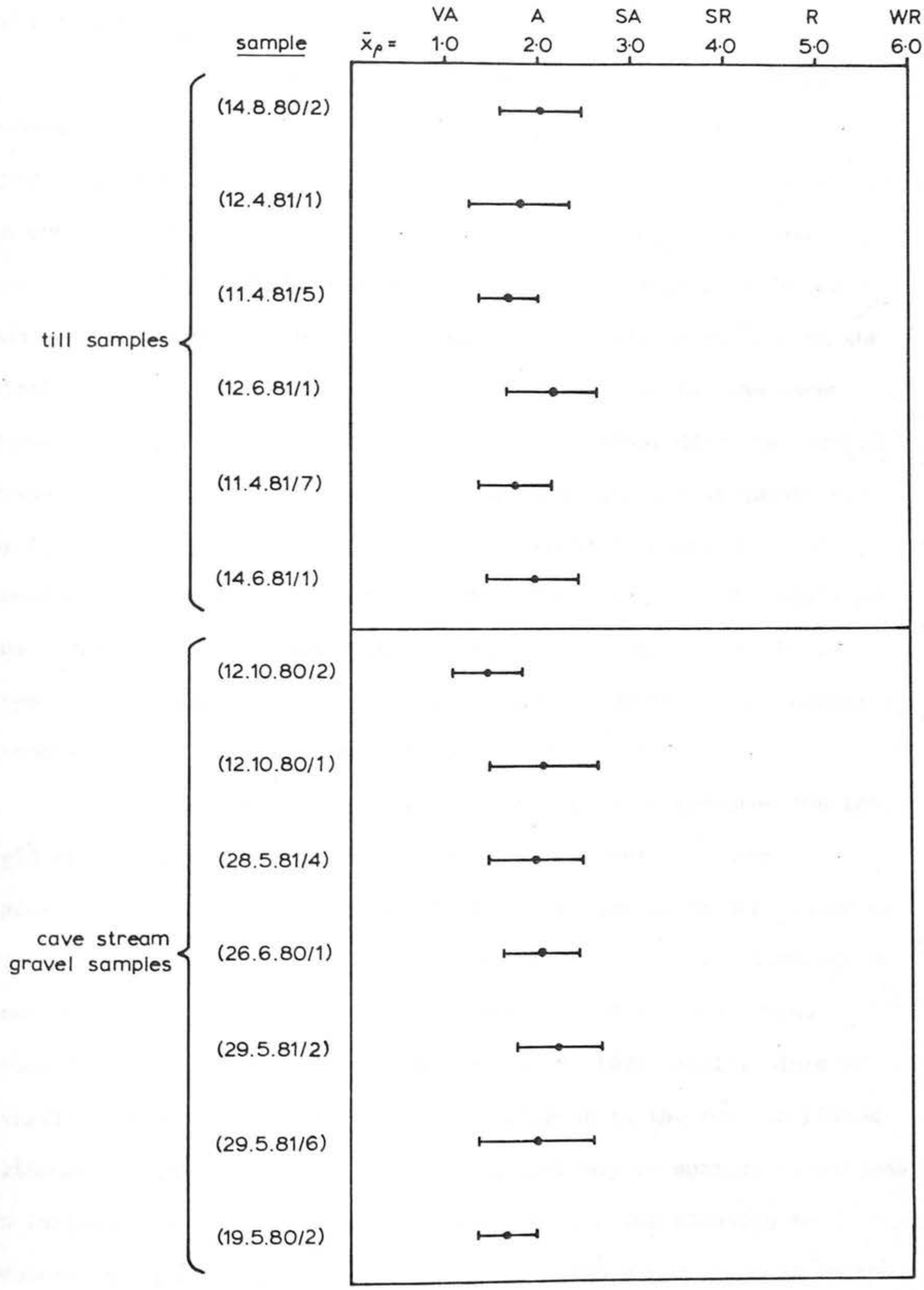
An attempt was made to quantify the relative proportions of autochthonous and allochthonous clasts in each cave gravel sample. Vosgesite intrusions occur within the dolomite bedrock and therefore stones of this lithology could be considered as either derived internally by breakdown or derived externally from glacial deposits. For the purpose of this study, half the vosgesite stones, together with all the dolomite clasts, were classed as autochthonous, and the remainder of the stones as allochthonous. The ratio of allochthonous to autochthonous particles was then computed ('D'-values in Fig. 10.3). Samples with low 'D'-values reflect large proportions of internally-derived stones. Excluding the three samples with high dolomite counts (Fig. 10.3(e), (f) and (g)), it can be seen that there is a large allochthonous component to the cave gravel samples. It is suggested that these allochthonous gravels were derived from local glacial deposits, washed into the caves by meltwaters on deglaciation and by floodwaters during interglacial periods.

Results of the roundness analysis of quartzite stones in the various cave gravel and till samples are shown in Fig. 10.5. There is little difference in the overall roundness values of the two sediment types, which both possess fairly angular quartzite clasts. The lack of appreciable rounding in samples from the caves can be accounted for by either low transportation distances or, more probably, the durability of this lithology preventing significant rounding in the fluvial environment from the angular forms attained in the preceding glacial environment.

Fig. 10.5 Roundness analysis of quartzite/Torridonian stones in certain Assynt cave gravels and the local till.

(N.b. the error bars represent $\pm$ 1σ from the mean.)

(diagram overleaf)



VA
A
SA
SR
R
WR
sample
$\bar{x}_p$ =
1·0
2·0
3·0
4·0
5·0
6·0
till samples
(14.8.80/2)
(12.4.81/1)
(11.4.81/5)
(12.6.81/1)
(11.4.81/7)
(14.6.81/1)
cave stream gravel samples
(12.10.80/2)
(12.10.80/1)
(28.5.81/4)
(26.6.80/1)
(29.5.81/2)
(29.5.81/6)
(19.5.80/2)

10.4 INTERPRETATION

10.4.1 Fine deposits

The literature on fine-grained deposition in caves has recently been reviewed by Bull (1981). Early work on sediments of this type, which are often laminated, concluded that they were deposited in ponded water due to climatically-induced flooding of the cave system. Sweeting (1950) ascribed a derivation from glacial deposits and sedimentation in underground, glacially-derived lakes, noting the similarity to varve deposition. The idea that these laminae were varves was subsequently developed (e.g. Siffre 1960, Masriera 1970). However, justification of the use of the term 'varve' was questioned by Bull (1976, 1977) as the laminations present in caves were not necessarily indicative of annual climatic fluctuations. He suggested the use of the term 'lamination' or 'rhythmite', each of which "is free of the underlying connotations of seasonal sedimentation normally associated with the term 'varve'" (Bull (1977, p. 87).

Bull's work in Agen Allwedd, S. Wales, questioned the long-held view (e.g. Bretz 1956) that laminated sediments in caves were deposited horizontally. He found evidence for laminated deposition on slopes over 70°, and suggested that a process of parallel accretion on underlying surfaces, first advocated by Reams (1968), would best account for his field observations (Bull 1976, 1977, 1981). This parallel accretion in flooded passages occurred to the roof in places. Although areally extensive, there were remarkably no spatial variations in laminae thickness or grain size from one sampling location to another, except in close proximity to joints and other fissures in the wall of the cave where laminae were slightly thicker. These cracks were suggested to be the routeways by which sediments reached the cave from the surface. The presence of laminated sediments was seen as the result of multi-source pulsed inputs of sediment-bearing water. Bull advocated a translatory flow mechanism "where a push-through effect of a water input into a column displaces water held in storage from an earlier infiltration occurrence. The water is then translated as a pulsed input and hence a sedimentation effect some distance below" (Bull 1981, p. 15). Alternate light and dark laminae were not only due to variations in grain size, but also variations in shape, with blocky particles forming the light layers and platey particles the dark ones. Persistence of similar laminae over large areas of the cave system, and supposedly in other caves in the region, was seen to suggest "climatic control rather than internally developed stochastic pulsations" (Bull 1981, p. 20).

In the Assynt caves, two suites of fine deposits have been identified, initially on the basis of their colour:

(i) 'Fossil cave silts'. Pale yellow silts and silty sands occupying the large, abandoned, phreatic passages in the larger cave systems.

(ii) 'Modern flood deposits'. Darker fine sands, silts and clays occupying areas of the caves prone to flooding.

Colour differences between the two groups are considered to reflect differences in sediment source.

The fossil cave silts are largely allochthonous. The dominant lithology is quartz, believed to be derived from the local quartzite-rich tills. The varying amounts of calcareous sediment probably represent the products of glacial abrasion of the dolomite bedrock. It is therefore suggested that these sediments represent the finer fractions of the surficial glacial deposits - the glacial 'rock flour' - washed into the caves by way of the various fissures running through the dolomite.

The relatively large proportion of clay-sized particles in the fossil cave silts (Fig. 10.2) suggests that they were deposited in still-water conditions. The areal extent of these deposits, often to the roof of passages, could only have been achieved in flooded cave systems. The presence of silt at levels higher than the present cave entrances requires those entrances to be blocked. In the absence of material evidence of these barriers, the only likely solution is that the cave entrances were blocked by glacier ice. It is therefore suggested that all the present evidence points to the deposition of these sediments under the last ice sheet.

There is an apparent lack of lamination in the fossil cave silts, except for those found in Uamh an Tartair (Traligill). The Uamh an Tartair section had been recently washed, revealing laminations that may exist in all the other dried-out sections. However, the pulsed input mechanism advocated by Bull (1981) in a periglacial situation may not necessarily hold true for sediments deposited in caves under glacier ice. Whether or not fine sediments were being washed into the cave in the latter case would be dependent on the presence or absence of basal meltwaters, in turn dependent on glaciological variables. It was suggested in chapter 6 that basal meltwaters were unlikely to be present at the glacial maximum, although striae indicate that temperate ice covered the area during the later phases of the last ice sheet. Deposition of much of the fossil cave silts may therefore also date from these later stages of the existence of the ice sheet. An essentially continuous input of sediment-laden water into flooded caves during this period would probably lead to mixing of particle sizes at the time of deposition, which might help to explain the apparent lack of lamination in many of the deposits.

The flood fines have also been deposited in quiet-water conditions, as water recedes after flooding events. Their darker colouration is thought to reflect a certain percentage of soil particles, washed in from the surrounding area, although a large proportion of their bulk is made up of particles derived from the local tills.

10.4.2 Stream gravels

There has been little written about stream gravels in caves. Siffre (1959) and Siffre & Siffre (1961) suggested that cave stream pebbles are highly rounded but more flattened than comparable pebbles from surface streams because of increased pressure flow. Similar results were presented by Sweeting (1972), but Bull (1976, 1978) questioned these conclusions as he found that gravel form, sphericity and roundness were controlled by rock type, and suggested that the previous studies might reflect a thinly-bedded cave breakdown deposit subsequently rounded and redeposited. Newson (1971), who included the Traligill valley as one of his study areas, concluded that the dominant movement of stream boulders and cobbles occurred during flood events, and then more often down the surface flood channels than through the cave systems. Bull's (1976, 1978) results appeared to support this view, with discrete sampling areas showing the effect of lithological dilution from local breakdown deposits that seemed to have largely remained in place since deposition.

In the Assynt caves, allochthonous and autochthonous components have been isolated in samples of some of the gravel deposits. A high allochthonous content in several of these samples and other gravel sections studied during fieldwork suggests a derivation from tills in the area. Local inputs of breakdown have resulted in samples showing varying amounts of included autochthonous material. The fluvial transportation of these gravel deposits is not in doubt :

many of the stones are rounded, and these deposits are clearly bedded in some of the studied sections. Evidence has been presented from Uamh an Tartair (Traligill) that the fossil cave silts were followed stratigraphically by gravel, deposited on the eroded surface of the silts and fining upwards into silts and clays. From the position of the section in the cave, this must have occurred prior to the establishment of the present (Postglacial) drainage routeways, and therefore the gravels probably represent meltwater drainage on deglaciation. Other stream gravels are associated with active streamways. In view of the work of Newson and Bull, these gravels must be considered only to be transported through the cave system by floodwaters when discharges are sufficiently large to carry them as part of the bedload.

10.4.3 Breakdown deposits

These are autochthonous deposits resulting from the collapse of cave roof and/or walls, and which have undergone little or no subsequent transportation. Davies (1949) suggested a genetic classification of block breakdown (due to the opening-up of joints), slab breakdown (due to the opening-up of bedding planes), plate breakdown (due to widening of closely-spaced joints) and quantitatively insignificant chip breakdown (due to release of compression within the bedrock). Davies (1951) extended this work to discuss the possible mechanics of the breakdown process, viewing the cave roof as both a fixed beam and a cantilever subjected to horizontal and vertical stresses. In the literature earthquakes, frost-shattering, solution along planes of weakness, undercutting of walls by cave streams, unloading after glaciation causing formation of dilatation cracks, and removal of internal support on drainage of the cave from phreatic conditions, have all been postulated as possible causes of breakdown (e.g. Warwick 1956, 1971; White & White 1969; Simons 1965; Miskovsky 1966; Tratman 1969; Sweeting 1972; Bull 1976).

Breakdown has been found in many locations throughout the Assynt caves and at a variety of stratigraphic levels (as the most recent deposits, situated on the surface of other cave sediments; above, below and within the fossil cave silts; and in various relationships with speleothem deposits). As the causes of breakdown are dependent on so many different factors, it is unlikely that one will find specific layers of breakdown that can be correlated from cave to cave across the study area, to be related to specific catastrophic events. Such breakdown events are usually only discrete phenomena occurring within different areas of various caves at different times. Furthermore, in the case of a single, thick deposit of large breakdown slabs or blocks, there are few ways of ascertaining whether or not one or several breakdown events occurred, as such deposits are unlikely to be subsequently moved in view of their size.

10.5 CONCLUSIONS

All the Assynt cave sediments have a large allochthonous content, except in certain localities where cave breakdown has diluted the lithological composition with dolomite fragments. It is suggested that a large proportion of the cave sediments was derived from local glacial deposits.

The composite clastic sediment stratigraphy of Uamh an Tartair (Traligill) can be proposed as a type sequence for at least a major part of the sedimentary history of the Assynt caves. Pale yellow silts and sandy silts are seen in various caves in the area and they are amongst the oldest clastic sediments present. They have been deposited variously on bedrock, breakdown or flowstone layers relating to earlier - possibly much earlier - phases in the geomorphic history of the respective cave. It is assumed that these fossil cave silts represent the same major depositional phase which can be correlated within the major cave systems. It has been suggested that they were deposited under the last ice sheet in flooded cave passages, sometime after the glacial maximum.

The next depositional phase was one of coarse fluvial sediments which are seen to overlie the yellow silts of the Uamh an Tartair sequence. The coarseness and form of some of these fluvial sediments indicates that stream discharges in the caves were high and turbulent. They have been attributed to the last deglaciation.

These sediments are post-dated in different caves by local breakdown deposits or speleothem formation. Postglacial vadose streams have selectively removed previous deposits, although the coarser sediments are only likely to be transported to new positions in the cave by extreme flooding events which afterwards leave many surfaces within the cave covered in dark brown fine deposits.

CHAPTER 11: GEOCHRONOMETRIC DATING OF THE ASSYNT KARST DRAINAGE SYSTEM

11.1 INTRODUCTION

Evidence has been presented in chapter 9 for a multiphase development of the major Assynt cave systems and progressive lowering of the local water-table (most clearly represented in the Traligill basin). The presence of certain cave sediments attributed to the time during which the last ice sheet existed implies that the caves containing them are older than the Late Devensian. In an attempt to find out how much older, recourse was made to chronometric dating techniques. Two radiocarbon dates from osseous material recovered from one of the Creag nan Uamh caves are discussed in chapter 12 : similar material in other Assynt caves is restricted to certain bones found on the surface of the deposits, and likely to be Postglacial in age. Other dateable materials of more widespread occurrence are speleothems (i.e. the various crystalline cave formations, those of calcite being the most common and the ones referred to below). Attempts have been made to date calcite speleothems by a variety of techniques (e.g. radiocarbon (Hendy 1970), palaeomagnetism (Latham 1977; Latham et al. 1979) and thermoluminescence (Wintle 1978)) but the most successful attempts so far have utilised the uranium decay series.

A number of speleothems were collected by the author, in collaboration with T.C. Atkinson, R.S. Harmon and P.L. Smart, from several caves in the Assynt area. Care was taken to note the location, stratigraphic position and potential significance with regard to cave development of each sample at the time of collection; samples were taken with a view to minimising the damage done to adjacent speleothem formations and where removal of a speleothem would not detract from the natural beauty of a flowstone display. The sampling strategy was to attempt to collect what appeared to be inactive speleothems from above, beneath and within the sedimentary infill of the caves, and from different situations that would give minimum and/or maximum dates for specific important events in the development of the underground drainage system and hence, by extrapolation, specific events in the evolution of the total drainage network of the particular basin. This was restricted by the limited distribution of suitable speleothems in the Assynt area and by the limitations of the dating method employed.

11.2 THE $^{230}Th/^{234}U$ DISEQUILIBRIUM DATING TECHNIQUE

This dating technique has recently been reviewed by Harmon *et al.* (1975), Gascoyne *et al.* (1978) and Schwarcz (1978, 1980). It is important to understand the limitations of all radiometric dating methods if the significance and reliability of the dates obtained are to be assessed; hence a simplified appraisal of the basic principles of this dating technique is presented here.

A radioactive decay series is the successive changes from an initial element (the 'parent nuclide') through various other radioactive isotopes (the 'daughter nuclides') to a stable element beyond which no further decay is possible. The decay series relevant to the dating of speleothems is the one by which uranium-238 decays by way of uranium-234, thorium-230, radium-226, lead-210 and other short-lived daughter nuclides to stable lead-206. If a natural material (*e.g.* rock) containing uranium is left undisturbed for a few million years, the degree of decay (per unit time per gramme of sample) of the daughter isotopes will equal the decay of the respective parent isotopes; a state of secular equilibrium is said to exist. On formation of a sedimentary deposit, various processes separate parent uranium isotopes from their daughter nuclides, initiating a state of disequilibrium. If a closed system is maintained and no post-depositional disturbance occurs, radioactive decay resumes at a steady rate until secular equilibrium is achieved once more. The date of deposition is found by measuring the extent to which secular equilibrium has been reached.

Ions of uranium and thorium are initially introduced into the environment from deep within the earth by volcanic activity. Subsequent weathering of igneous rocks releases these ions. Thorium is insoluble and its high valency state means that it is quickly absorbed onto clay particles, but uranium forms various soluble complexes after oxidation (mainly uranyl carbonates and phosphates) and is readily removed by groundwaters. The groundwaters eventually reach the sea where the uranium is subsequently incorporated into carbonate sediments by chemical precipitation. Thorium is also introduced into the marine sediments on clay particles washed into the sea. The sediments later consolidate to form rock which may subsequently be uplifted to form limestones on the earth's surface.

Solution of this limestone once more releases the uranium and thorium ions. The thorium is again absorbed onto clay particles, but uranium as soluble uranyl carbonate can percolate down through the limestone with the groundwaters. On entering a cave space, CO_2 — charged groundwaters may become supersaturated due to degassing of CO_2, which causes the uranyl carbonate to dissociate to UO_2^{2+} and CO_2. The uranyl ion UO_2^{2+} is co-precipitated with calcite as speleothems, forming a theoretically thorium-free system; it should be noted, however, that thorium on clay particles may have entered the cave with the drip-waters and been incorporated into the flowstone. As it can be assumed that, in a closed system, the thorium-230 present has been produced solely by the radioactive decay of uranium-234 and uranium-238, the age of the speleothem can be calculated by measuring the amount of thorium-230 present in the sample in relation to the amount of uranium-234, taking into account the initial disequilibrium between uranium-234 and uranium-238. It is possible to check for the presence of contaminating thorium, i.e. thorium on clay particles incorporated into the speleothem. Contaminating thorium-230 is usually found in association with another daughter nuclide, thorium-232. It has been suggested (R.S. Harmon, pers. comm.) that a sample with a $^{230}Th/^{232}Th$ ratio of less than 10 shows unacceptable thorium contamination and hence the age obtained for that sample is unreliable.

Uranium-234 has a half-life of c. 248,000 years, compared with the half-life of carbon-14 of c. 5,700 years : therefore uranium-234 is said to have a low 'activity rate'. Uranium concentrations in speleothems are very low, ranging from 0.01-100.0 p.p.m., though normally 0.1-2.0 p.p.m. The low activity rate and low concentrations of uranium mean that the standard deviations given with $^{230}Th/^{234}U$ dates tend to be greater than those for radiocarbon dates, but the former's extended potential dating range of between 1,000 and 400,000 years B.P. makes it a powerful dating method especially beyond the effective range of the radiocarbon technique (i.e. c. 40,000 years B.P., unless isotopic enrichment methods are used when the upper limit is extended to 60,000-70,000 years B.P.).

Four essential conditions must be met if a sample is to be dated by the $^{230}Th/^{234}U$ method :

(i) The sample should contain sufficient uranium for analysis (greater than 0.05 p.p.m.).

(ii) Samples should not contain more than 1% insoluble detritus.

(iii) A sample should not show signs of recrystallisation or partial dissolution.

(iv) A series of dates along a sample should be in correct stratigraphic order (i.e. increasing in age from top to bottom of a stalagmite and vice versa for a stalactite).

11.3 RESULTS

At the time of writing, 18 age determinations had been made on 13 samples collected from Uamh an Claonaite in the Allt nan Uamh basin and five caves in the Traligill basin : these results are presented in Table 11.1. Other samples still await analysis (due to the delay in setting-up a new dating laboratory in the University of East Anglia after the closure of the one at the Scottish Universities Research and Reactor Centre, East Kilbride).

The dates obtained so far range from 2,000 $\pm$ 1,000 years B.P. (SU6-80) to 192,000 $^{+\,53,000}_{-\,39,000}$ (810530-6). Only one sample (SU4-80) showed appreciable thorium contamination ($^{230}Th/^{232}Th$ ratio = 7), and therefore this date must be viewed as being too old by something of the order of a thousand years. Caution is also advised when considering the date of 11,000 $\pm$ 2,000 years B.P. from sample SU16-80 as a certain amount of thorium contamination seems likely ($^{230}Th/^{232}Th$ ratio = 17).

Table 11.1 Uranium concentrations, isotope activity ratios and $^{230}Th/^{234}U$ ages for Assynt speleothems

Sample No.	Description	Cave	U conc. (p.p.m.)	$^{234}U/^{238}U$	$^{230}Th/^{234}U$	$^{230}Th/^{232}Th$	Age 10^3 years B.P.)
SU16-80	flowstone	Uamh Cailliche Peireag	0.07	1.32 ± 0.04	0.09 ± 0.02	17	11 ± 2
SU4-80	stalactite	Uamh an Tartair (Traligill)	0.23	1.29 ± 0.04	0.11 ± 0.01	7	12 ± 1
SU11-80	stalagmite (top)	Uamh an Tartair (Traligill)	0.43	1.19 ± 0.03	0.05 ± 0.01	> 200	5 ± 1)
	(mid)	" " " "	0.28	1.16 ± 0.03	0.07 ± 0.01	> 200	7 ± 1)
	(base)	" " " "	0.30	1.10 ± 0.03	0.08 ± 0.01	> 200	9 ± 1)
SU12-80	stalagmite	Uamh an Tartair (Traligill)	0.27	1.17 ± 0.04	0.21 ± 0.02	34	26 ± 2
SU13-80	stalagmite	Uamh an Tartair (Traligill)	0.22	1.25 ± 0.03	0.054 ± 0.009	> 200	5.5 ± 0.8
SU19-81	stalagmite (base)	Lower Traligill Cave	0.18	1.12 ± 0.06	0.41 ± 0.06	155	$56 {+14 \atop -12}$
SU6-80	stalagmite	Tree Hole Cave	0.20	1.04 ± 0.03	0.02 ± 0.01	> 200	2 ± 1
SU1-80	eroded flowstone (top)	Firehose Cave	0.12	1.29 ± 0.06	0.21 ± 0.02	30	26 ± 3)
	(base)	" "	0.13	1.21 ± 0.03	0.28 ± 0.04	> 200	38 ± 6)
SU2-80	stalagmite	Firehose Cave	0.18	1.13 ± 0.03	0.08 ± 0.02	> 200	9 ± 2
AU1-80	flowstone (top)	Uamh an Claonaite	0.10	1.50 ± 0.06	0.46 ± 0.03	64	63 ± 6)
	(base)	" " "	0.09	1.52 ± 0.05	0.60 ± 0.04	> 200	122 ± 12)
AU2-80	stalagmite (top)	Uamh an Claonaite	0.08	2.32 ± 0.05	0.24 ± 0.03	31	30 ± 4
810530-3	flowstone (top)	Uamh an Claonaite	0.10	1.42 ± 0.04	0.86 ± 0.05	18	$181 {+24 \atop -18}$
810530-6	flowstone (top)	Uamh an Claonaite	0.13	1.40 ± 0.05	0.77 ± 0.04	39	$143 {+16 \atop -13}$)
	(base)	" " "	0.13	1.57 ± 0.08	0.89 ± 0.08	> 200	$192 {+53 \atop -39}$)

11.4 SPELEOTHEM DATES AND THE CHRONOLOGY OF THE GEOMORPHIC EVOLUTION OF THE ASSYNT KARST DRAINAGE NETWORKS

As speleothems cannot be formed beneath water, all the dates obtained post-date the lowering of the local water-tables, when the originally phreatically-formed passages became abandoned or changed to conditions of vadose water flow. However, from the results of the sedimentological analyses (Chapter 10), it has been suggested that all the caves in the Assynt district became flooded during a glaciation (presumably the last) when glacier ice covered the dolomite area, blocking cave entrances and prohibiting the free drainage of subglacial meltwaters. In view of this it is evident that, since their initial phreatic formation, the cave systems have probably been filled with water on a number of occasions in the past, during the many glaciations that have doubtless affected the Assynt area. Hence it is probable that calcium carbonate formations collected from a particular area of a cave are representative of several suites of speleothems of differing ages (assuming that earlier speleothems survive subsequent inundation). The older speleothems in such cases are liable to possess features of resolution and hiatuses in their growth structure: therefore it is necessary to take the form of a speleothem into consideration before categorically stating that the date obtained from it post-dates the last drainage of the water-filled cavern where it was located.

Ten of the dates so far obtained from the Assynt speleothems pre-date the last full glacial period as currently defined by the British radiocarbon chronology (*i.e.* *c.* 26,000 - *c.* 14,000 years B.P.) (*e.g.* Atkinson *et al.*, in prep.). From the locations of the dated speleothems it is interesting to consider to what extent the present cave systems had developed before this phase.

The five oldest dates were obtained from Uamh an Claonaite. Blocks of broken flowstone, showing at least two thick, laminated calcite layers separated by a thin, mud-stained hiatus, were collected from the breakdown chamber between sumps 2 and 3 at the lower end of Mud Passage (Fig. 9.10). Sample AU1-80 yielded a date of 122,000 $\pm$ 12,000 years B.P., which equates well with dates for oxygen-isotope substage 5e of the marine record (Shackleton & Opdyke 1973, 1976; Shackleton & Matthews 1977) correlated with the last interglacial period as defined by terrestrial evidence (Shackleton 1969; Shackleton & Heusser 1977; Mangerud et al. 1979); this is the first Scottish radiometric date from the last interglacial (Lawson 1981a). The top of the same sample yielded a date of 63,000 $\pm$ 6,000 years B.P. Three older dates have since been obtained : sample 810530-6 yielded dates of 143,000 $^{+16,000}_{-13,000}$ (top) and 192,000 $^{+53,000}_{-39,000}$ (base) years B.P.; sample 810530-3 yielded a date of 181,000 $^{+24,000}_{-18,000}$ (top) years B.P. Hence the high-level, abandoned passages of Uamh an Claonaite, and also those of Allt nan Uamh Stream Cave and the Creag nan Uamh caves which are thought to relate to the same early subterranean drainage system, probably pre-date the penultimate interglacial period. The top of sample AU2-80, a stalagmite situated on the downstream side of Sump 1 near the present streamway (Fig. 9.10), indicated that this part of the active drainage route in Uamh an Claonaite was in existence by 30,000 $\pm$ 4,000 years B.P.; in view of its large size, the passageway was probably formed long before this date.

In the Traligill basin, the Cnoc nan Uamh cave system is the only one to possess high-level, formerly phreatic passageways, but these are almost completely choked with silts and sands and are devoid of speleothems. The oldest date from this complex cave is one of

26,000 $\pm$ 2,000 years B.P. (SU12-80) from a stalagmite at the entrance to Beechbarrow Chamber (Fig. 9.2). Samples from The Grotto, which possesses the best speleothem displays in the cave, indicate that this part of the system was in existence at least 9,000 years B.P. A date of 12,000 $\pm$ 1,000 years B.P. was obtained from a stalactite (SU4-80), but this date is unreliable due to the presence of contaminating thorium (section 11.3). Nevertheless, older speleothems probably exist in The Grotto, which is at approximately the same altitude as Beechbarrow Chamber; indeed, it is expected that age determinations predating the last glacial period will be obtained from as yet undated speleothem samples collected in May, 1980, from positions above and within the gravel fill. The Glenbain Hole - Firehose Cave drainage system is a fairly mature one, as indicated by dates of 26,000 $\pm$ 3,000 (top) and 38,000 $\pm$ 6,000 (base) years B.P. from a piece of eroded flowstone (SU1-80). The same can be said of Lower Traligill Cave, where the base of a stalagmite was dated to 56,000 $^{+14,000}_{-12,000}$ years B.P. (SU19-81). This date indicates that the Traligill Main Thrust was in operation as a main aquifer in the basin by this time.

The $^{230}Th/^{234}U$ dates obtained so far from the Assynt area show that most, if not all, of the elements of the Allt nan Uamh and Traligill drainage systems existed prior to the Late Devensian glaciation. During the Late Devensian the only major effect on the drainage networks seems to have been a general filling of the caves with clastic sediments. The only dated speleothem sample directly overlying such sediments (thus giving a minimum age for their deposition) is flowstone overlying gravel in Uamh Cailliche Peireag : a date of 11,000 $\pm$ 2,000 years B.P. (SU16-80) is consistent with deposition of the gravel by meltwaters on deglaciation. Several other samples from similar stratigraphic positions, as yet undated, are expected to provide similar results. The removal of bedrock from the main valleys by glacial erosion has meant that, on deglaciation, pre-glacial water-tables have been lowered, resulting in drainage of the previously flooded caves and vadose incision into the sediments and bedrock. The utilisation and development of open rifts and fissures that were previously located in the phreas but now form part of the active streamways probably occurred at this stage.

The broken blocks of flowstone from Uamh an Claonaite which have been dated give maximum ages for the breakdown events that emplaced them. Breakdown occurred sometime after 63,000 $\pm$ 6,000 years B.P., sometime after 143,000 $^{+16,000}_{-13,000}$ years B.P. and sometime after 181,000 $^{+24,000}_{-18,000}$ years B.P.

The speleothem dates indicate that parts of the Assynt cave systems are of some considerable age, and that a majority of the elements of the present underground drainage network were in existence before the development of the last ice sheet. The hypothesis of certain past workers (Ford 1959; Warwick 1962, p. 182; Jeffries 1972, p. vii) that the present caves of Assynt are young features and predominantly Postglacial in age is therefore clearly untenable.

11.5 WIDER IMPLICATIONS OF THE ASSYNT SPELEOTHEM DATES

Speleothems are only deposited in caves if certain special conditions exist. If the exact conditions restricting the growth of speleothems can be isolated, the clustering of speleothem dates should indicate periods during which these limiting conditions were not operating. For speleothems to form, groundwaters containing enough CO_2 to be able to dissolve $CaCO_3$ in the limestone must be available. These groundwaters must then enter a cavern space whose atmospheric conditions allow the reprecipitation of $CaCO_3$. Speleothems are deposited primarily by degassing of CO_2 (when the partial pressure of the CO_2 contained in the percolating groundwaters is greater than that of the cave atmosphere) and consequent supersaturation of the solution with calcite. In certain circumstances, evaporation of the solution has the same effect (*e.g.* White 1976).

Therefore, two main conditions seem to be necessary for the deposition of speleothems:

(i) a supply of water;

(ii) a reservoir of CO_2-rich gas from which the groundwaters may acquire a higher partial pressure of CO_2 than the cave air.

Such conditions exist in British caves at present, however "there are good reasons for supposing that speleothem deposition ceased, or was at least drastically reduced, during past periods of periglacial and glaciated (*sic*) conditions" (Atkinson *et al*. 1978, p. 25). Work in the Canadian Arctic (Smith 1972; Cogley 1972) has shown that the potential aggressiveness of karst groundwaters is low despite the theoretical increasing solubility of CO_2 in water with decreasing temperatures (Corbel 1959a, 1959b). This has been attributed to the poorly developed vegetation cover in the study areas, for it is clear that biogenic CO_2 is the largest reservoir of that gas available to groundwaters (Smith 1972; Woo & Marsh 1976). CO_2 in water is driven off on freezing, hence meltwaters from snow and glacier ice do not contain enough CO_2 for degassing in caves. This, together with the fact that percolation of groundwaters would almost totally cease in areas of continuous permafrost, suggests that speleothem deposition would be prevented or substantially decreased during glaciation or periglaciation of an area.

Recently, attempts have been made to isolate periods of speleothem deposition, and hence non-glacial/non-periglacial periods, from an analysis of the temporal distribution of speleothem dates. A study of 54 speleothem dates from caves in alpine North America (Harmon _et al_. 1977) indicated clustering into five distinct groups:

(i) 15,000 years ago to the present day;

(ii) 90,000 - 150,000 years ago;

(iii) 185,000 - 235,000 years ago;

(iv) 275,000 - 320,000 years ago;

(v) earlier phases before 350,000 years ago.

Frequency histograms of these dates (_ibid_., Fig. 2) show a certain agreement between the phases of speleothem deposition and the peaks of the oxygen-isotope curve from deep-sea core V28-238 (Shackleton & Opdyke 1973). Analysis of the distribution of 28 speleothem age determinations enabled Atkinson _et al_. (1978) to isolate four periods when speleothem deposition occurred in British caves. The four periods were:

(i) 17,000 years B.P. to the present day (_i.e_. Lateglacial and Flandrian periods);

(ii) around 60,000 years B.P. (? Chelford Interstadial);

(iii) 90,000 - 140,000 years B.P. (oxygen-isotope stage 5 of the marine record);

(iv) earlier depositional episodes of uncertain duration before 170,000 years B.P..

Gascoyne (1981) reported similar results from the dating of speleothems from caves in NW England. These speleothem growth periods agree well with those determined from N America (Thompson _et al_. 1974, 1976; Harmon _et al_. 1975).

The 18 speleothem dates from the Assynt caves indicate similar depositional phases to those found by Atkinson _et al_. (1978), with the addition of speleothem growth between 26,000 and 38,000 years

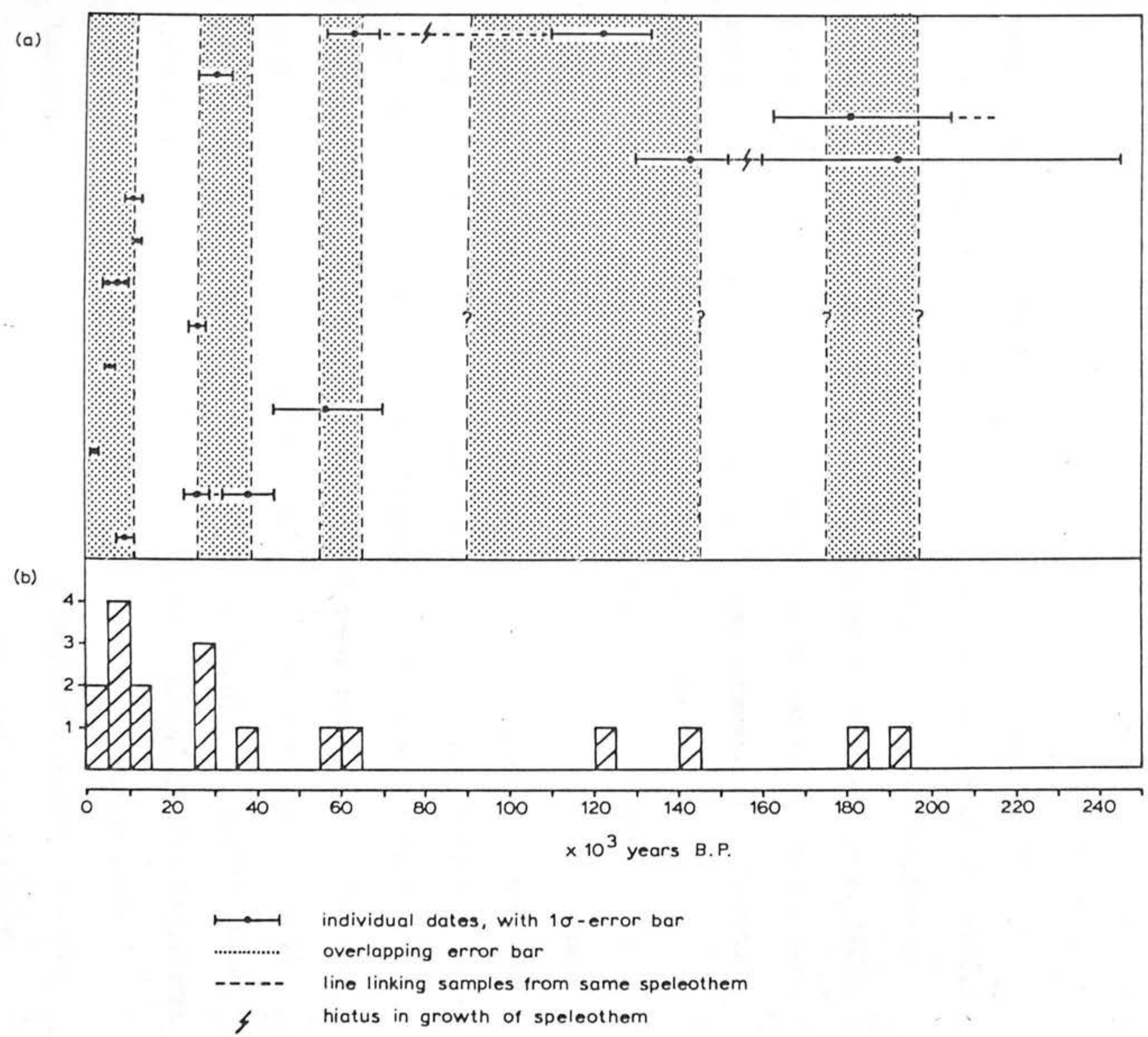
B.P. and a reduction of the most recent phase to one that occurred between *c*. 11,000 years B.P. and the present day (Fig. 11.1). Sissons (1981) has recently criticised the radiocarbon chronology of the Scottish Late Quaternary period, which is largely based on a few equivocal dates. It is interesting to note that the Assynt speleothem dates are consistent with the Scottish radiocarbon chronology as well as supporting the Late Quaternary chronology of Britain as a whole (Atkinson *et al*., in prep.).

Fig. 11.1 Speleothem dates from the Assynt area.

(a) Temporal distribution of speleothem dates, showing minimum growth periods (shaded areas) based largely on results obtained by Atkinson _et al._ (1978). (_N.b._ the length of suggested growth periods for the oldest Assynt speleothems so far dated are necessarily very speculative, being based on so few dates.)

(b) Frequency histogram of age determinations from the Assynt area _per_ 5 ka periods.

(diagram overleaf)


(a)
(b)
4
3
2
1
?
0
20
40
60
80
100
120
140
160
180
200
220
240
x 10^3 years B.P.
individual dates, with 1σ-error bar
overlapping error bar
line linking samples from same speleothem
hiatus in growth of speleothem

CHAPTER 12 : THE CREAG NAN UAMH BONE CAVES

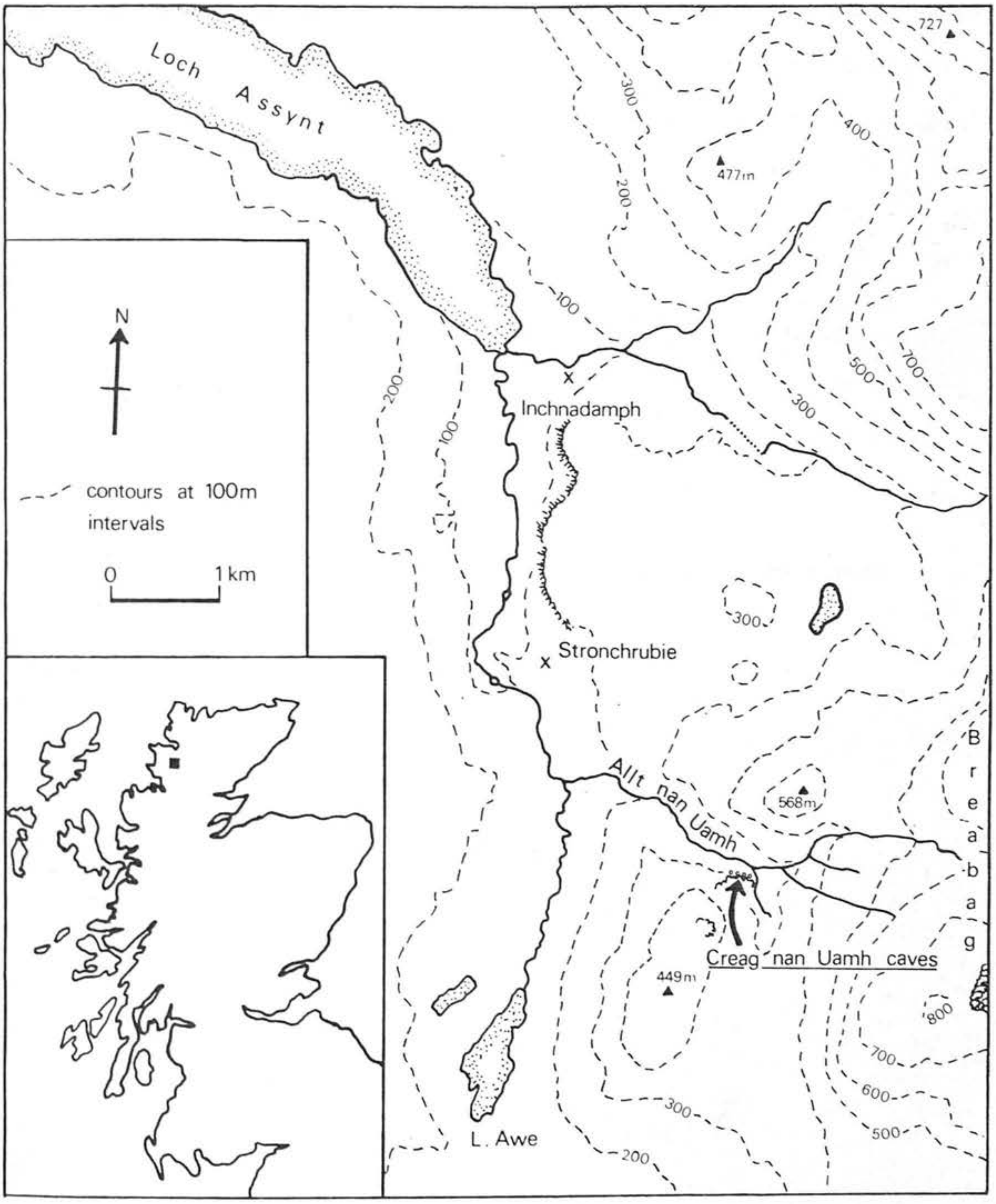
12.1 INTRODUCTION

Reference was made in chapter 9 to the Creag nan Uamh caves (Fig. 12.1). In view of their palaeoenvironmental (and archaeological) significance they will be considered in greater detail in this chapter. The Creag nan Uamh contains three caves of fairly large dimensions, and a number of niches and rockshelters (Fig. 12.2). The three main caves have been excavated on two occasions in the past, but the significance of the results of these excavations is somewhat enigmatic due mainly to poor documentation of the work.

B.N. Peach and J. Horne excavated the easternmost of the three caves - Bone Cave - in 1889 (Peach & Horne 1892b, 1917). Their findings remained unchallenged until 1926 and 1927, when the entrance chambers of all three caves were systematically excavated by J.E. Cree, J.G. Callander and J. Ritchie. Despite the fact that preliminary reports of the first fieldwork season of these later excavations were published (Callander *et al*. 1927; Cree 1927; Ritchie 1928), a final report was never produced. Although the archaeological significance of the sites was commented on, often somewhat disparagingly, in several standard textbooks of the period (*e.g*. Childe 1935; Movius 1942; Lacaille 1954), the caves were gradually forgotten, except for a few references in the caving literature (*e.g*. Ford 1959) where little attempt was made to add to the preliminary reports. Much information regarding the 1926-1927 excavations has been preserved in the Royal Scottish Museum, Edinburgh, in the form of unpublished letters and

Fig. 12.1 Location of the Creag nan Uamh caves.

(diagram overleaf)



Loch Assynt
727
300
400
477m
200
100
700
500
300
200
100
N
contours at 100m intervals
0
1 km
Inchnadamph
Stronchrubie
300
Allt nan Uamh
568m
Breabag
Creag nan Uamh caves
449m
800
700
600
500
300
200
L. Awe

a

b

manuscripts. The present writer has recently examined this archive material and published a paper discussing the results of these excavations (Lawson 1981b).

This chapter will concentrate on the Creag nan Uamh caves as palaeoenvironmental and archaeological sites, and discuss the geomorphic significance of the deposits they contain. An outline of previous work in the caves will be followed by an examination of the author's present analyses of the cave sediments and faunal remains.

12.2 PAST WORK IN THE CAVES

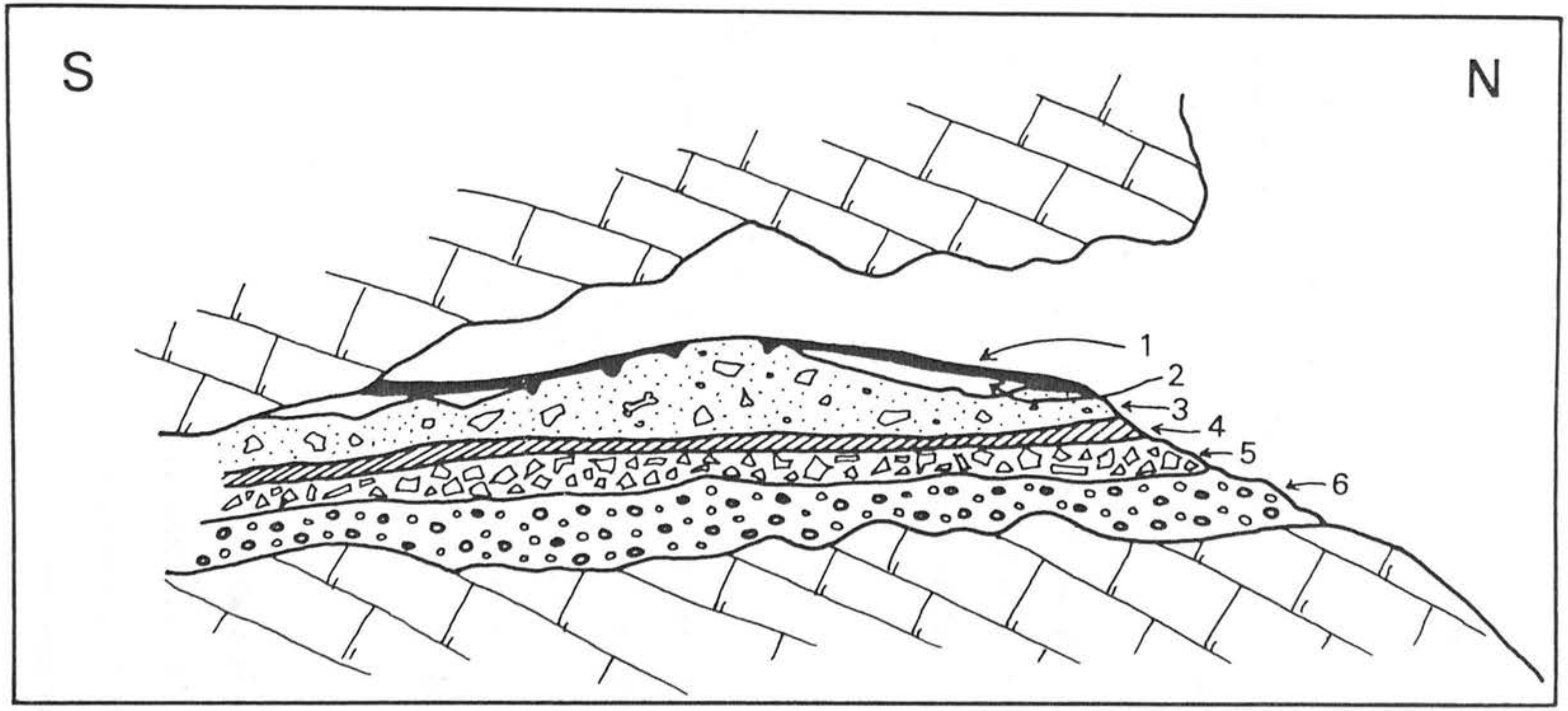
12.2.1 The 1889 excavation in Bone Cave

The six-layered stratigraphy described by Peach & Horne (1892b, 1917) is reproduced schematically in Fig. 12.3. The lowest stratum (layer 6) was composed of well-rounded stones containing many lithologies foreign to the dolomite area (e.g. quartzites and porphyrites). Imbrication in the gravels was said to indicate their deposition in a stream issuing from the mouth of the cave. The gravels were covered by a layer of fine splinters of the local dolomite (layer 5), interpreted as the products of frost action on the roof and walls of the cave. An arctic fauna was recovered from this layer. Layer 4 was a compact grey clay containing quartzite pebbles, said to resemble the glacial deposits in the adjoining valley : this layer was interpreted as having been derived from the quartzite slopes of Breabag, "... carried on the surface of a glacier and shot (sic) into the cave from the lobe of ice that passed down the valley" (Peach & Horne 1917, p. 341). Serious doubt is cast on this rather peculiar mode of deposition by the fact that all the tills in the vicinity tend to have a yellowy-

Fig. 12.3 Schematic cross-section through the deposits of Bone Cave (redrawn from Peach & Horne (1917)).

(N.b. for stratigraphy, see text.)

(diagram overleaf)


S
N
1
2
3
4
5
6

orange matrix, and examples of a local grey till have not been encountered during the course of the present research. Layer 3 was described as "a true cave earth" (ibid., p. 341), consisting of a red clay containing many limestone fragments and faunal remains. The presence of man in the cave at the time this layer was accumulating was said to have been indicated by "fireplaces, split and burnt bones ..., (although) no artefacts were detected" (ibid., p. 341). This layer was surmounted by a layer of whitish calcareous marl (layer 2) and a final layer of sheep dung.

Peach & Horne (1917, p. 327) were of the opinion that the early history of Bone Cave related to the "late glacial time, or at least to a period before the final disappearance of local glaciers in that region".

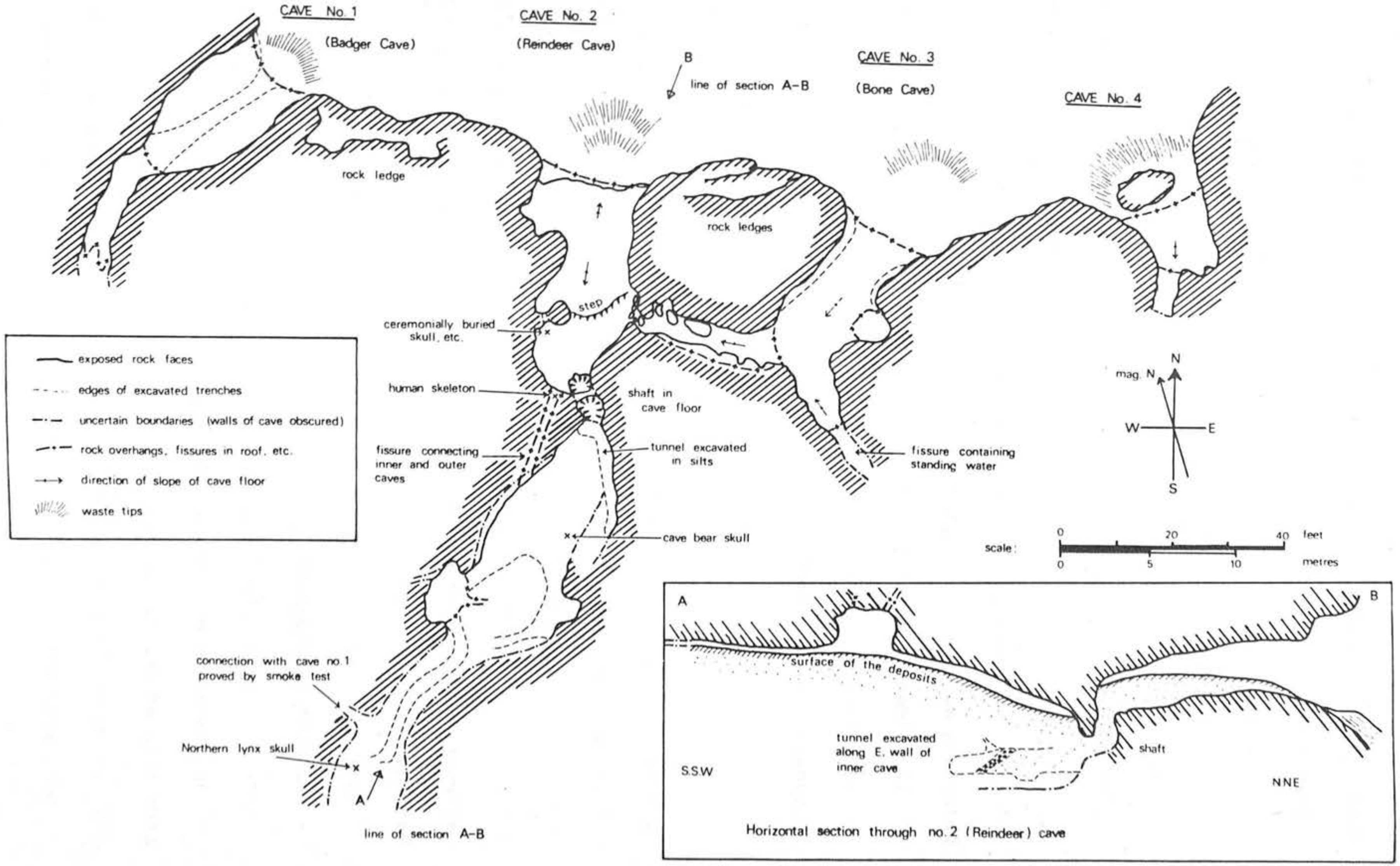
12.2.2 The 1926-1927 excavations

Amongst the previously unpublished archive material pertaining to the later set of excavations was found a plan of the caves, surveyed in 1927 by a Mr. McWilliams. Fig. 12.4 is based on McWilliams's survey. Details of the excavations are presented in Lawson (1981b) and hence only an outline of the main findings will be given here.

(a) Badger Cave Previously referred to only as Cave no.1 (Callander et al. 1927; Ford 1959; Lawson 1981b), it is suggested that this cave should be called Badger Cave after the large quantity of the remains of these animals found here during the course of various excavations (section 12.4). The cave comprises two chambers leading to a series of crawl-passages at the back, which are largely blocked by silts and

Fig. 12.4 Plan of the Creag nan Uamh caves (redrawn from the unpublished 1927 survey by McWilliams).

(diagram overleaf)


CAVE No. 1
(Badger Cave)
CAVE No. 2
(Reindeer Cave)
B
line of section A–B
CAVE No. 3
(Bone Cave)
CAVE No. 4
rock ledge
rock ledges
step
ceremonially buried skull, etc.
human skeleton
shaft in cave floor
fissure connecting inner and outer caves
tunnel excavated in silts
fissure containing standing water
cave bear skull
connection with cave no. 1 proved by smoke test
Northern lynx skull
A
line of section A–B
exposed rock faces
edges of excavated trenches
uncertain boundaries (walls of cave obscured)
rock overhangs, fissures in roof, etc.
direction of slope of cave floor
waste tips
mag. N
N
W
E
S
scale:
0
20
40
feet
0
5
10
metres
A
B
surface of the deposits
tunnel excavated along E. wall of inner cave
shaft
S.S.W
NNE
Horizontal section through no. 2 (Reindeer) cave

breakdown blocks (Fig. 12.4).

The stratigraphy of the outer chamber consisted of an upper red 'cave earth' up to 0.9 m deep, more consolidated towards the base, overlying a layer of 0.6 - 1.2 m of bluish clay containing angular limestone fragments. The inner cave was filled with about 1.2 m of reddish-brown silt that had little or no stratification. Since no water-worn pebbles were found, no active streamway seems to have been present in the cave throughout its depositional history.

All the faunal remains were found in the red 'cave earth', especially in its upper part. A complete faunal list will be presented in section 12.4. Archaeological finds were sparse and probably all came from the upper layer of the outer chamber. They included hearths and charcoal flakes, and several other items of dubious significance (Lawson 1981b).

(b) Reindeer Cave This is the largest cave on the crag (Fig. 12.4), consisting of an outer chamber and a large inner cave, connected by a 2.7 m-deep shaft. Total excavation of the outer chamber has left none of the original deposits, but the discovery amongst the archive material of two of Cree's drawings of sections through the deposits has greatly aided reconstruction of the stratigraphy; these drawings are reproduced as Figs. 12.5(a) and (b). Fig. 12.5(a) shows the profile 2.4 m from the entrance and can be taken as typifying the stratigraphy from the mouth of the cave back to the rock step (Fig. 12.4). From this step to the back of the outer chamber, the stratigraphy was as shown in Fig. 12.5(b), the section 4.6 m from the entrance. It can be seen that an additional gravel layer was found towards the back of the cave. This stratum was interpreted as having been introduced into the cave from Bone Cave by way of the connecting passageway (Fig. 12.4).

Fig. 12.5 The stratigraphy of Reindeer Cave in 1927, from drawings by J.E. Cree (presented in Lawson, 1981b).

Key: (a) Cross-section 2.4 m from the entrance:

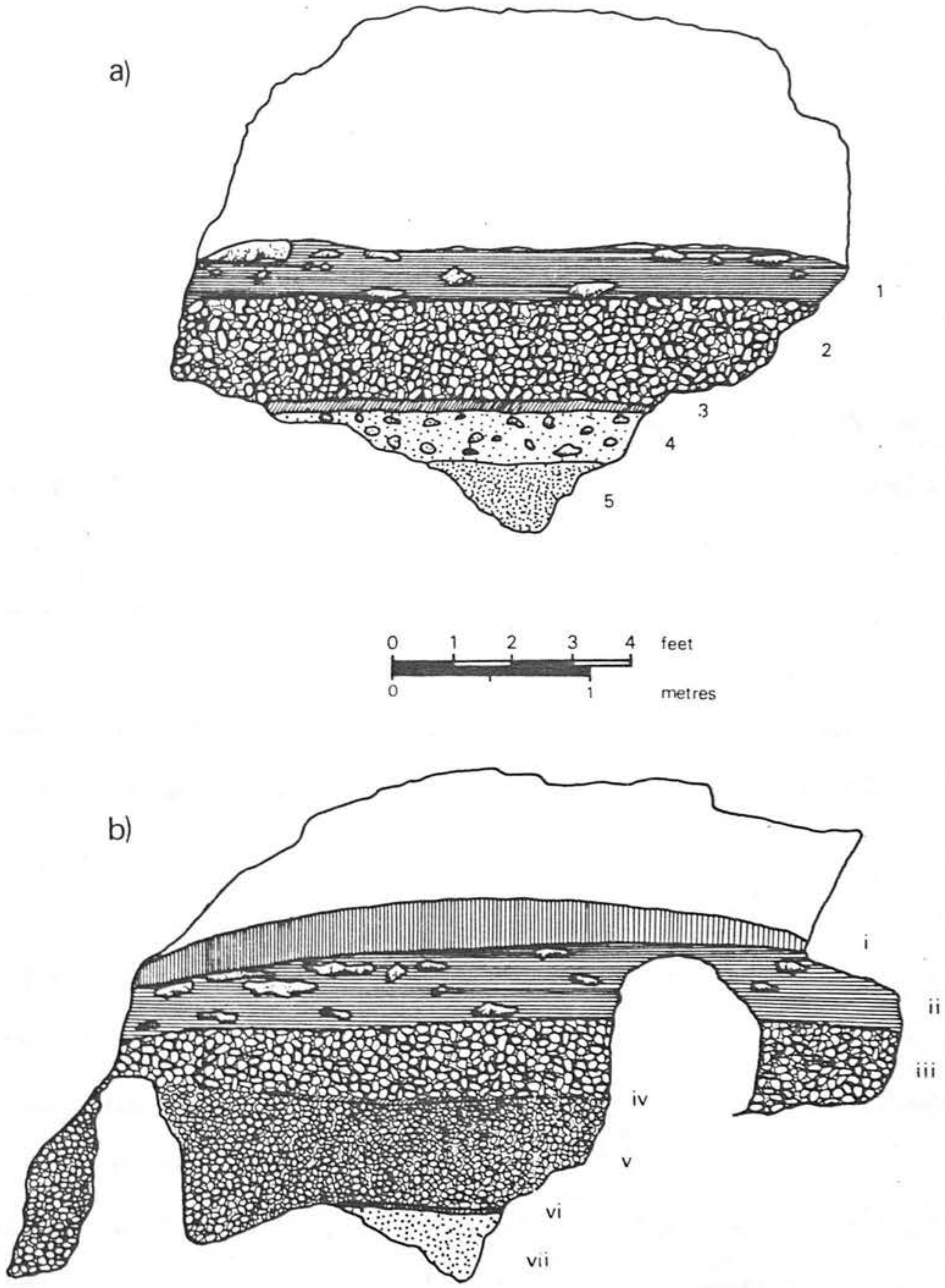
1. 'Cave earth' with fallen roof-stones.
2. Subangular gravel containing the bones of reindeer, *etc.*
3. Grey sand.
4. Grey clay containing quartzite.
5. Greyish-yellow clay.

(b) Cross-section 4.6 m from the entrance:

i. Peaty material (sheep dung).
ii. 'Cave earth' with fallen roof-stones.
iii. Subangular gravel containing the bones of reindeer, *etc.*
iv. Grey sand.
v. Barren gravel.
vi. Greyish sand.
vii. Yellow clay.

(diagram overleaf)

a)



b)

Certain sedimentary characteristics of the principal layers within the stratigraphy were gleaned from an unpublished report written by J. Phemister of the Geological Survey (Lawson 1981b).

(i) The lower gravel layer (Fig. 12.5(b), layer v) was visibly different from the upper gravel (Fig. 12.5(b), layer iii), being devoid of faunal remains and containing a greater proportion of quartzite, vosgesite, felsite and porphyrite pebbles. The clasts also showed a greater degree of roundness.

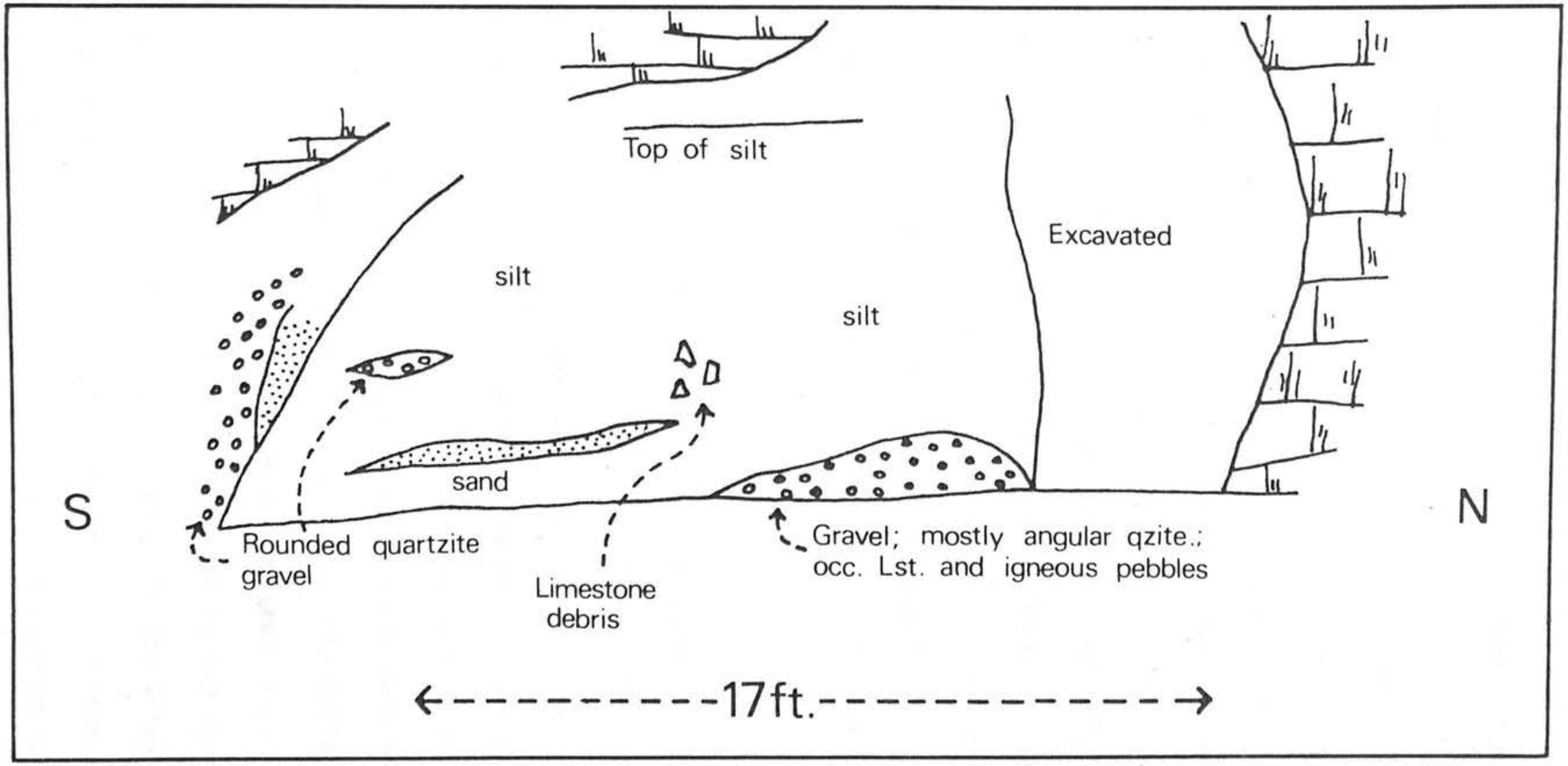
(ii) The upper gravel was composed largely of limestone fragments, generally subangular in appearance, but pebbles of quartzite, vosgesite, felsite and porphyrite were also reported. Layers of sand and grit were apparently found within this stratum, but no details were given. This layer contained numerous remains of reindeer and other animals.

Both these layers were inclined at an angle of about 15° towards the back of the entrance chamber of the cave, according to Phemister's report. The fissure at the rear of the chamber contained some of the upper gravel layer and the overlying layers of 'cave earth' and sheep dung. The shaft connecting the inner and outer caves was full of gravel and sand : the presence of many reindeer antlers and bones amongst the gravel suggests that it relates to the upper gravel unit of the outer chamber, but no definite statement to that effect was made by the excavators. A discrete layer of limestone splinters was found at a depth of 1.8 - 2.1 m, and at 2.7 m depth the gravel gave way to pale yellow silts.

These silts are the same ones that almost fill the inner cave to its roof. Trial trenches were dug over and through part of these deposits (Fig. 12.4), which have remained largely intact since then. Fig. 12.6 reproduces Phemister's drawing of the section revealed

Fig. 12.6 Section through the deposits of the inner chamber of Reindeer Cave, as seen by J. Phemister in 1927. Section exposed on western side of trench dug southwards from the bottom of the shaft into the inner chamber. (Redrawn from Phemister's unpublished report, after Lawson (1981b).)

(diagram overleaf)


S
N
Top of silt
Excavated
silt
silt
sand
Rounded quartzite gravel
Limestone debris
Gravel; mostly angular qzite.; occ. Lst. and igneous pebbles
17 ft.

by a trench dug southwards from the bottom of the shaft. Discrete layers and pockets of sands and gravels are shown in the silts, indicating a complicated depositional history. The steeply dipping bed of water-worn gravels shown on the left-hand side of Fig. 12.6 was analysed for lithological composition by Phemister. A count of 50 pebbles gave the following results : quartzite 50%, vogesite 26%, felsite 14%, limestone 10%. This contrasts with the results of a similar count of pebbles from the outer cave : limestone 48%, quartzite 18%, Pipe Rock 2%, vosgesite 30%, diorite 2%. Faunal remains were found in the upper gravel unit and the overlying 'cave earth'. A discussion of these finds and a faunal list are presented in section 12.4; consideration of the character and significance of the exceptionally large quantity of pieces of reindeer antler will be left till then. A few scanty finds of possible archaeological significance have been discussed in Lawson (1981b).

(c) Bone Cave A reassessment of Peach and Horne's original excavations in Bone Cave (Fig. 12.4) was undertaken in 1927 in order to correlate the stratigraphy with that of Reindeer Cave. At an early date, the excavators doubted the effectiveness and quality of the work of Peach and Horne (Lawson 1981b), posing the problem of how much credence to give to the stratigraphy of the cave published by these authors (1892b, 1917). The stratigraphy in Bone Cave as seen by Callander and Cree was as follows. The basal deposit was a dark grey clay, containing particles said to be silicified limestone, but more likely to be chert or quartz fragments which occur as discrete bands in the local dolomite. A layer of clay up to 0.6 m deep, containing quartzite stones, overlaid this stratum; it had a very irregular surface and varied in colour from a very yellow clay on the west side of the cave to quite a red colour in the centre, possibly due to differences in the amount of roof-drip. Above this, on the west side of the cave only, was found a clean grey gravel. It was in a discrete channel which ran from the entrance of the cave, through the connecting side passage (hereafter referred to as Connecting Passage) to Reindeer Cave, where it formed the lower of the two gravel units. Overlying the gravel in the tunnel and western side of the cave, and above the clay layer elsewhere, was found red 'cave earth', much disturbed by burrowing animals. The 'cave earth' in the tunnel was capped by a thin flowstone layer. Faunal remains are detailed in section 12.4. No archaeological remains were found in this cave.

12.2.3 Present work in the caves

Particular attention was paid during the present study to determining the amount of in situ cave sediments in the Creag nan Uamh caves, in order to check some of the sedimentary characteristics and modes of deposition postulated by the previous excavators. Unfortunately nearly all the deposits in the outer chambers were removed during the 1926 and 1927 excavations. A trial trench dug across Badger Cave, approximately 5 m from the cave entrance, showed that the apparent sedimentary fill comprises material that has slumped from the degraded sides of the median trench dug in 1926. The dolomite bedrock surface immediately below these disturbed deposits was very weathered and in certain places had decayed completely to leave a grey silty material containing insoluble particles of chert and quartz. As the dolomite is so easily weathered, it is reasonable to assume that the bluish clay with limestone fragments reported from the base of the deposits in Badger Cave, together with the grey clay containing chert particles at the base of the Bone Cave stratigraphy, represent similarly weathered bedrock.

The only *in situ* deposits in the outer chamber of Badger Cave are a series of yellow silts and sands occupying rock ledges on the western side (Fig. 12.7). The only visible sedimentary features are load structures. No features indicative of a unidirectional water flow are apparent. These silts and sands were not commented on by the previous excavators. Much of the original sediments of the inner chamber remains. Friable red 'cave earth', containing numerous small bones of various amphibians, occupies ledges around the cave walls; in places it overlies flowstone deposits or is heavily indurated with a calcium carbonate cement. The remainder of the inner chamber is floored with a yellow-brown silty-sand, which may be the equivalent of the reddish-brown silt described by the previous excavators. It contains many faunal remains in its upper layers, but has been greatly disturbed by rabbits and badgers whose remains are abundant.

No *in situ* deposits remain in the entrance chamber of Reindeer Cave or in the fissure and shaft at the rear. However, the deposits previously described for the inner cave appear to have been largely untouched since the 1927 excavation. A number of different strata can be seen at different sections in the cave, where the deposits have been dissected by the trenches of the excavators. At no point in the cave was the bedrock floor reached by trial pits; it is quite possible that these deposits extend considerably farther than can be seen at present (both laterally and vertically), but without specialist equipment it would have been hazardous to proceed deeper. The lowest stratum visible is a silty-clay containing fallen roof-stones that have completely weathered away to a grey clay. A lens of manganese-stained gravel, 80-150 mm thick, separates this layer from reddish-yellow silts (330 mm deep) containing a few fallen roof-stones. A concentration of these angular dolomite cobbles separates this layer from 450 mm of pale

Fig. 12.7 Schematic section through the deposits in the entrance chamber of Badger Cave.

Key: 1. Silt.

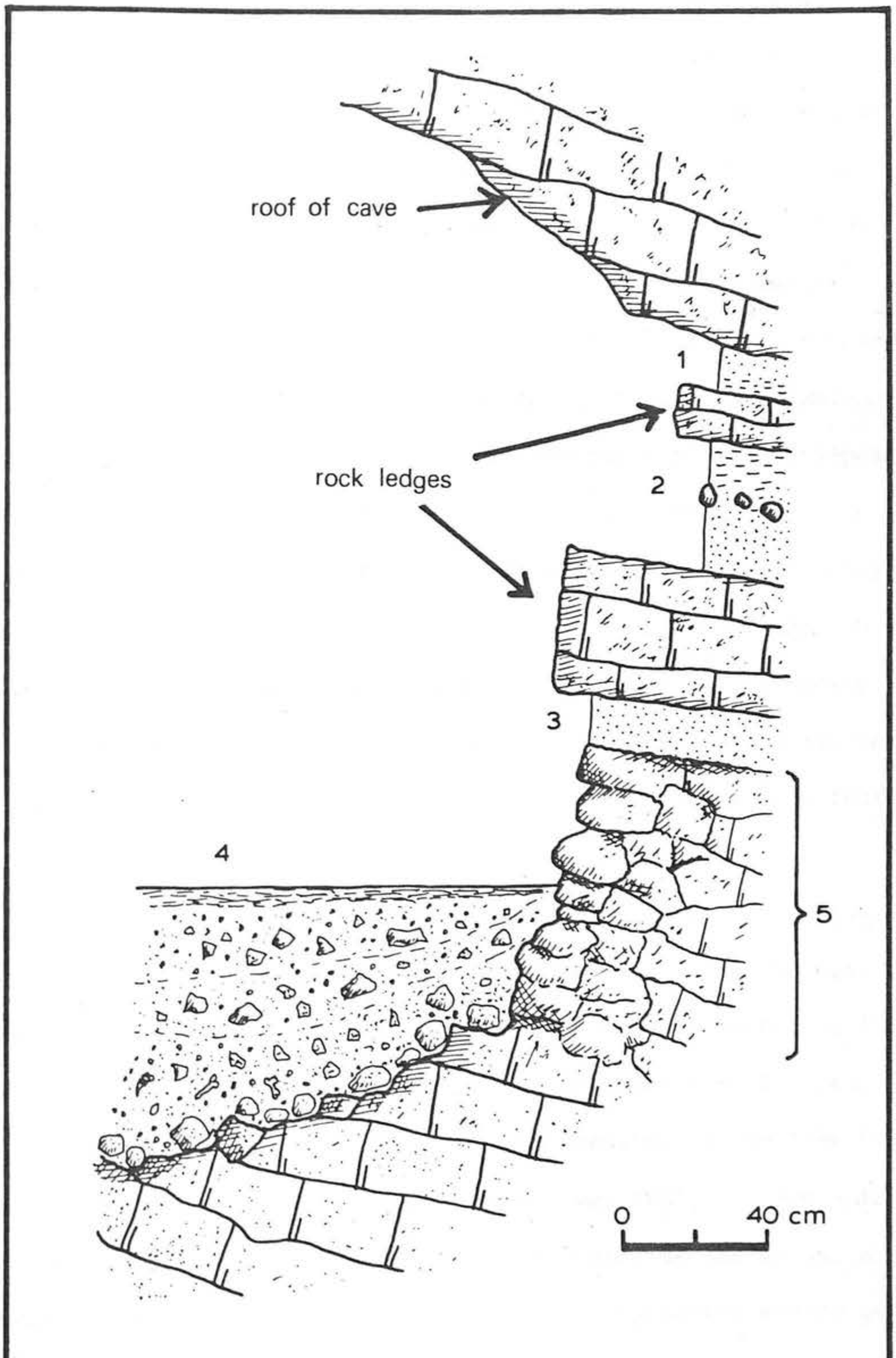
2. Silty-sand containing breakdown fragments.

3. Silty-sand.

4. Mixed deposits, slumped from degraded sides of the trench excavated in 1926.

5. Weathered dolomite bedrock.

(diagram overleaf)



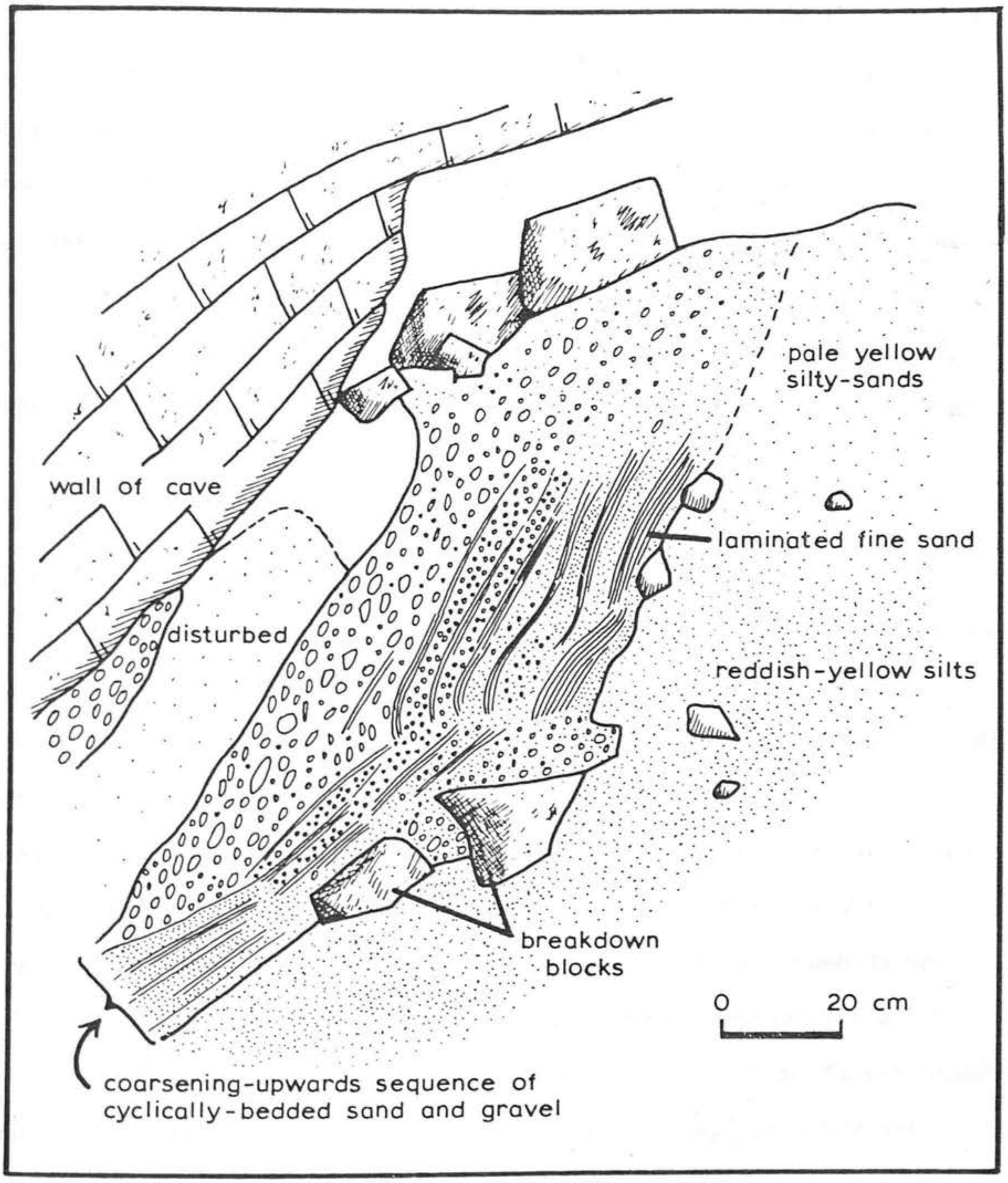
roof of cave
1
rock ledges
2
3
4
5
0
40 cm

yellow silts. The whole profile is topped with slabs of dolomite breakdown. On the eastern side of the chamber, and shown in profile at the southern end of the trench extending along the eastern wall from the foot of the shaft, a cyclically-bedded deposit can be seen cutting through the other deposits, coarsening upwards from laminated medium and fine sands, through coarse sands to rounded pebbles in a silty matrix (Fig 12.8). The gradually curving cave wall and roof act as the upper surface of this layer. This stratum is the same as that depicted in Phemister's sketch (Fig. 12.6). The long axes of rounded pebbles tend to dip steeply downwards, along the line of the laminae. The gravels appear to cut through all the previously mentioned deposits, save for the breakdown blocks on the very top. In several locations the pale yellow silts abut against the roof of the cave. Preliminary attempts to determine the relationship between the silts of the inner cave and the gravels formerly occupying the shaft failed, primarily because excessive digging would have deepened the shaft which was the sole means of access to the inner cave and an already hazardous feature to negotiate.

Few of the original deposits of Bone Cave exist _in situ_. A good section through stream gravels occurs in Connecting Passage. Cyclically bedded, these deposits are not quite as well sorted as those described above in the inner Reindeer Cave. The presence of a rock bar at the eastern end of Connecting Passage has resulted in the formation of foreset bedding on the leeward side (_i.e._ away from the cave entrance). Coarse gravel and cobbles are overlain by laminated medium sands, a fining-upwards sequence of gravel, topped by poorly-sorted coarse gravel and cobbles. The whole profile is approximately 1 m thick. These gravels can only represent those described in 1927 as also forming the lower gravel stratum in Reindeer Cave. Underneath the gravels, a

Fig. 12.8 Sketch of the section exposed at the end of the trench dug southwards from the bottom of the shaft, in the inner chamber of Reindeer Cave.

(diagram overleaf)



pale yellow silty-sands
wall of cave
laminated fine sand
disturbed
reddish-yellow silts
breakdown blocks
0 20 cm
coarsening-upwards sequence of cyclically-bedded sand and gravel

yellow-brown sticky mud was discovered filling pockets in the floor of the passage. A thin layer of flowstone (3 mm thick) caps the overlying gravels. In the small fissure containing standing water at the rear of Bone Cave on the eastern side (Fig. 12.4), a brown mud occurs, containing small bones and overlain by a thin layer of flowstone. This silty deposit is dissimilar from the 'cave earths' present in the Badger Cave, and it is therefore believed to equate with the upper layers of the so-called 'clay' described here in 1927, underlying the red 'cave earth'. At the very entrance to the cave, a cemented breccia can be seen at floor level. It is impossible to say what its relationship is to the original sedimentary fill of Bone Cave, as no evidence of the adjoining layers has been preserved.

12.2.4 Synthesis and the isolation of problems

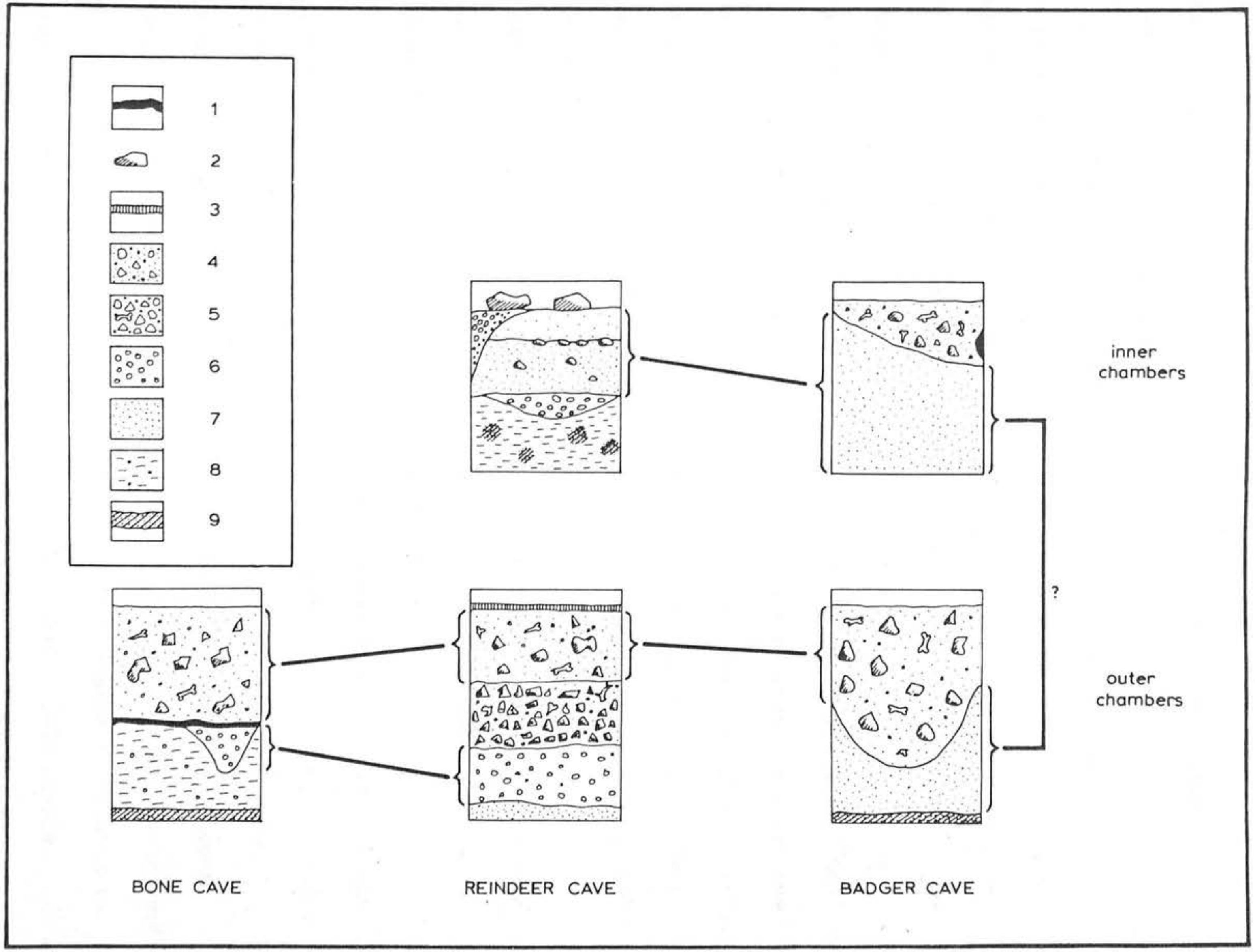
Fig. 12.9 is an attempt to reconstruct the basic stratigraphy of all the Creag nan Uamh caves. Correlations between strata are indicated where possible. The relationship of the sequence of silts and sands to the 'cave earth' in the outer chamber of Badger Cave is entirely speculative; it assumes a relationship between these silts and sands and the yellow-brown silts of the inner cave, which is not unreasonable although impossible to prove due to the disturbed nature of the latter. A relationship also possibly exists between the silts of the inner recesses of Badger Cave and one or more of the fine-grained deposits in the inner Reindeer Cave; although the two caves are not linked by any penetrable passageways, a smoke test conducted in August 1926 proved that the two caves are in fact connected (Lawson 1981b).

Having ascertained the basic stratigraphy of the caves, several problems were isolated that required further analysis of the deposits and the remains they contained. The presence of fluvial gravels

Fig. 12.9 Diagrammatic reconstruction of the lithostratigraphy of the Creag nan Uamh caves, showing postulated relationships between certain of the layers.

Key:
1. Speleothem deposit.
2. Breakdown material.
3. Organic layer of sheep dung.
4. 'Cave earth'.
5. Fossiliferous 'upper gravel' unit.
6. Water-lain gravels.
7. Silts and sands.
8. Clayey silts.
9. Weathered dolomite.

(diagram overleaf)



1
2
3
4
5
6
7
8
9
inner chambers
?
outer chambers
BONE CAVE
REINDEER CAVE
BADGER CAVE

in a cave high above present streams, and the presence of fine deposits in the inner chambers of the caves indicating complete deposition to the roof, are just two facets of an obviously complicated depositional history that must be explained.

Phemister believed that both gravel layers in the outer chamber of Reindeer Cave were deposited by a stream which entered the mouth of the cave and flowed down the shaft at the back. The gravels found in the inner chamber were not deposited by the same stream. He suggested that the silts were deposited earlier in the flooded inner chamber by a glacial stream that entered the cave from the back. The presence of gravel rather than silt in the shaft between the outer and inner chambers was, in Phemister's view, due to subsidence of the latter when the gravel-bearing stream entered the cave (Lawson 1981b). Cree proposed a similar depositional history for Reindeer Cave, except the introduction of the lower gravel from Bone Cave was preferred to Phemister's idea of the stream actually entering the mouth of Reindeer Cave. Both explanations, however, fail to clearly suggest where this stream flowed to after it had entered the shaft.

A further problem was the determination of the ages of the different strata in order to integrate them into the geomorphic history of the study area. The presence of ossiferous material obviously provided the opportunity for radiometric dating of some of the upper layers in the caves, and it was believed that evidence of the environments of deposition from the results of the sedimentary analyses, related to other geomorphic evidence from the surrounding area, would enable suggestions to be made as to the possible significance of the lower strata. Finally, it was hoped that an examination of the faunal content of the different lithostratigraphic units would provide information about the environment outside the cave at the time these layers were deposited.

12.3 SEDIMENTOLOGICAL ANALYSIS

12.3.1 Introduction

The techniques used to analyse the sediments from the Creag nan Uamh caves are the same as those described in section 10.2. Analysis of a number of samples collected in 1926 and 1927, and hitherto stored in the Royal Scottish Museum, was supplemented by analysis of samples collected during the present examination of the caves. For the purpose of this study, samples from the old excavations were given prefixes of 'RC' (= Reindeer Cave) and 'BC' (= Bone Cave); other samples, collected by the writer, have sample numbers based on the date of collection. Sample descriptions are given in Appendix I.

Section 12.2 has shown that the most complete stratigraphy of all the Creag nan Uamh Caves (*i.e.* the one containing most of the lithostratigraphic units of the composite profile) is from Reindeer Cave. Unfortunately, *in situ* sediments in the outer chamber of this cave are lacking, thereby reducing the amount of information that could be collected from a sedimentary sequence (*e.g.* presence of certain sedimentary structures) and necessitating almost complete reliance on the subjective assessment of the previous excavators. In order to reduce this subjectivity, it was decided to analyse one particular set of samples collected in 1926 and which appeared to represent a systematic sampling of the outer Reindeer Cave profile at approximately 30 cm intervals, regardless of the excavators' formal subdivision of the stratigraphy. It was hoped that a detailed sedimentary characterisation of these samples would enable the reconstruction of a quasi-'type section' for the outer chamber of Reindeer Cave which could be integrated into a composite lithostratigraphy of the Creag nan Uamh caves. As will be shown, this was only partially successful.

12.3.2 Results

(a) Analysis of samples from the outer chamber of Reindeer Cave

A primary consideration was to determine whether the samples from the outer Reindeer Cave section separated on sedimentological grounds into the three major lithostratigraphic units identified by the excavators (*i.e.* 'cave earth', upper gravel and lower gravel). The results of particle size analysis and lithological analysis of these samples are presented in Fig. 12.10.

Particle size analyses investigated both the total (or 'global') distribution of each sample and the distribution of the size fraction less than -2ϕ; both distributions are shown for each sample in Fig. 12.10. All the samples were polymodal. As particle size distributions reflect percentage weights in the different size fractions rather than percentage frequencies of grains of similar size, the total sample distributions tended to show large modes in the coarsest size fraction, where a few large particles may represent up to 55% of the total sample weight. More attention was therefore paid to the particle size distributions of the less than -2ϕ fraction (*i.e.* diameters less than 4 mm), also shown in Fig. 12.10. Modes were present at -1ϕ and 2ϕ (except for RC8), and weaker modes at 7ϕ for samples RC1 to RC5 and 6ϕ for samples RC6 to RC9. Sample RC1 had a strong mode at 5ϕ, otherwise there was a dearth of particles of this diameter. On the basis of particle size analysis of the less than -2ϕ fraction, it therefore seems that a tripartite division of the sediment profile is

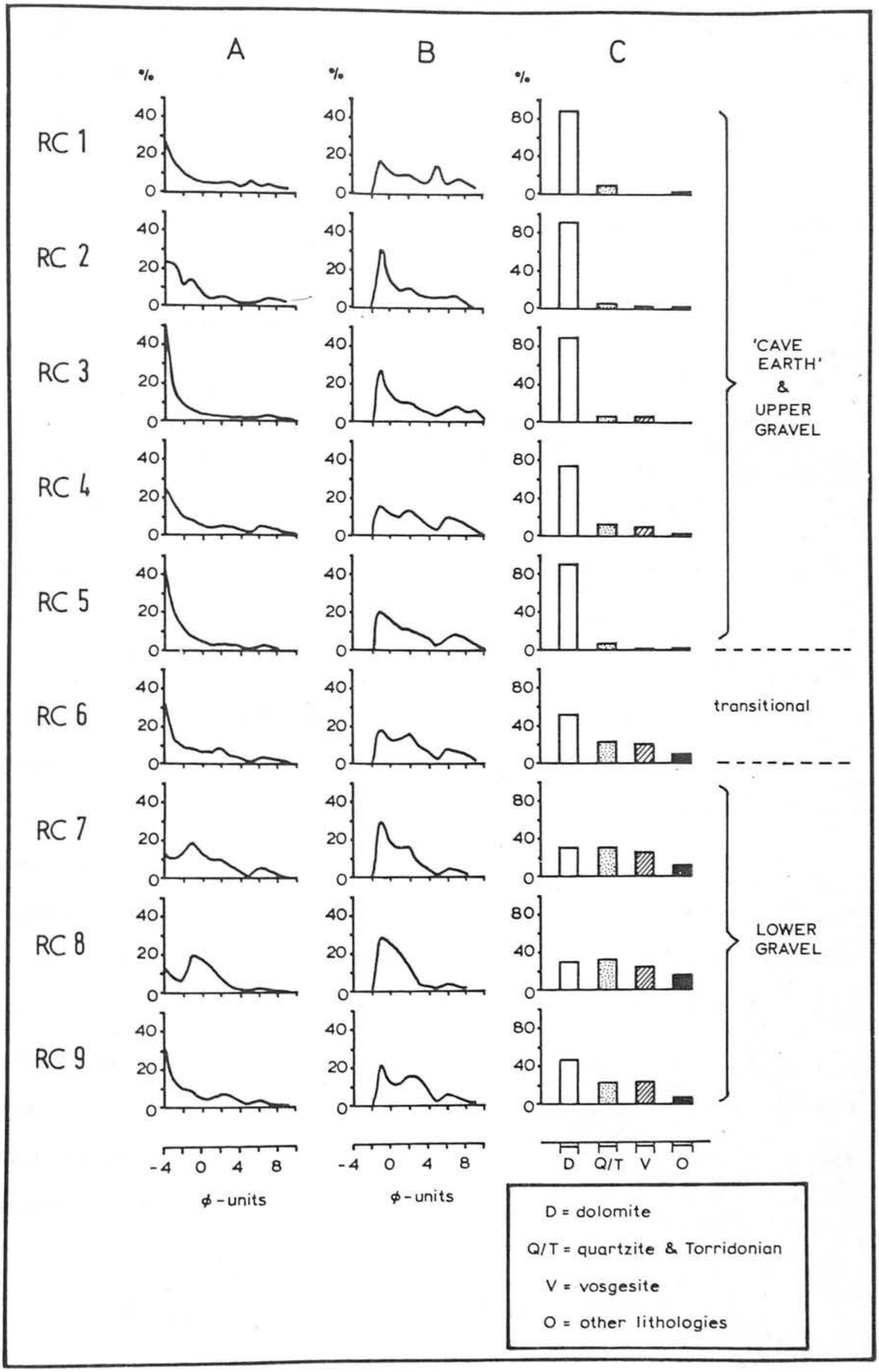
Fig. 12.10 Sedimentological analysis of the outer Reindeer Cave profile.

Key: A. Global particle size distribution.

B. Particle size distribution of the less than -2ϕ fraction.

C. Lithological composition of the -2ϕ to -3ϕ fraction.

(diagram overleaf)


A
B
C
%
RC 1
RC 2
RC 3
RC 4
RC 5
RC 6
RC 7
RC 8
RC 9
40
20
0
80
-4 0 4 8
ϕ - units
D Q/T V O
'CAVE EARTH' & UPPER GRAVEL
transitional
LOWER GRAVEL
D = dolomite
Q/T = quartzite & Torridonian
V = vosgesite
O = other lithologies

possible: sample RC1 comprised coarse and medium sands and coarse silts; samples RC2 to RC5 comprised coarse and medium sands in varying proportions with medium-fine silts; samples RC6 to RC9 were composed of coarse and medium sands (except RC8 which is unimodal about -1ø) with medium silts.

Results of the lithological analysis of the 4-8 mm diameter (-2ø to -3ø) size fraction are presented in Fig. 12.10. An attempt to divide the Reindeer Cave profile on statistical grounds by the chi-square test of difference between the lithological composition of subjacent samples was only partially successful. No significant difference at the 0.05 significance level was found between samples RC2 and RC3 (χ^2 = 0.49 with 2 d.f.) and samples RC7 and RC8 (χ^2 = 5.79 with 3 d.f.). A large degree of dissimilarity existed between samples RC5 and RC6 (χ^2 = 131.79 with 3 d.f.). Elsewhere differences occurred between subjacent samples. In samples RC1 to RC5, such differences can be explained by varying amounts of non-calcareous lithologies as the dolomite pebble fraction dominates these samples (Fig. 12.10), ranging from 73.3% to 90.1% of the total sample. Lower down the profile, quartzite, Torridonian sandstone, vosgesite and other lithologies, especially Breabag porphyrite, occurred in large quantities. Variations between the overall lithological content of samples RC1 to RC5 and RC7 to RC9 are clearly apparent from Fig. 12.10, with sample RC6 as transitional. It is suggested that the division of the profile at sample RC6 reflects the differences between the upper and lower gravels. No division between the upper gravel and the overlying 'cave earth' was possible on lithological grounds.

(b) The relationships between the various gravel layers

The incidence of several gravel units has been reported from various parts of the cave system (i.e. the upper and lower gravel layers which no longer exist, gravels in Bone Cave still present in Connecting Passage, and the gravels of the inner Reindeer Cave section). Samples from Bone Cave and the inner Reindeer Cave were compared lithologically with the upper and lower gravel units of the Reindeer Cave outer chamber by chi-square tests.

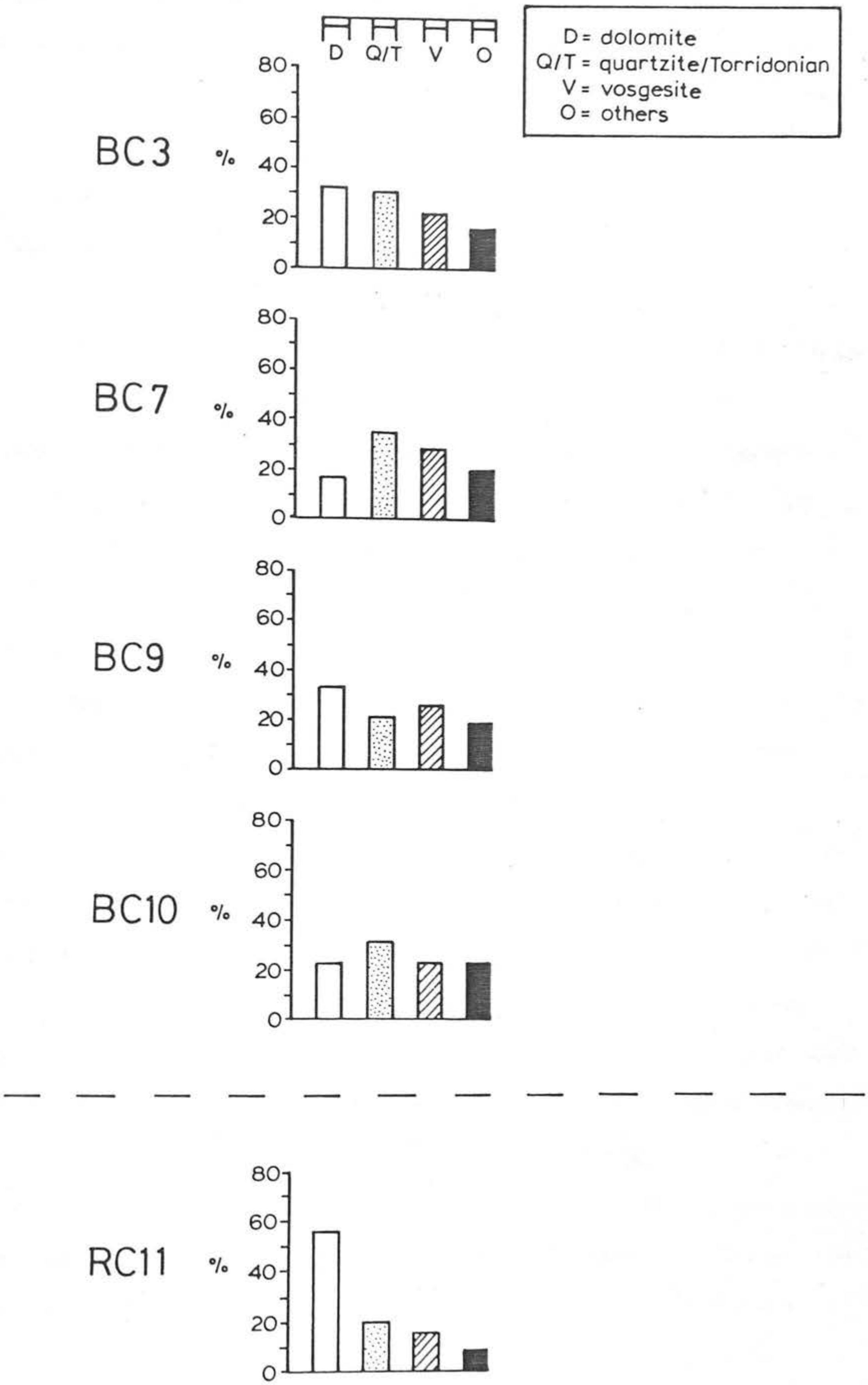
In view of the lack of strong statistical similarities between samples of the lower gravel from Reindeer Cave, it was not surprising to find that four available samples of the Bone cave gravel (BC3, BC7, BC9 and BC10) showed few statistical similarities to either the former or amongst themselves. No significant difference at the 0.05 significance level was found between samples BC3 and RC7 ($\chi^2 = 2.07$ with 3 d.f.) and at the 0.01 significance level between samples BC7 and BC10 ($\chi^2 = 8.52$ with 3 d.f.); otherwise samples were lithologically different, statistically. Nevertheless, Fig. 12.11 shows that the lithological composition of these samples includes high proportions of quartzite and vosgesite, and therefore they are more akin to the lower gravel samples from Reindeer Cave than the upper (dolomite-rich) gravel. This, plus the fact that the Bone Cave gravel was traced in 1927 through Connecting Passage to form the lower gravel in Reindeer Cave, would seem to equate the two sets of samples.

Two samples of the gravel from the inner Reindeer Cave (RC11 and 6.8.79/2) were distinctly different from the other gravel samples, possessing a higher percentage of dolomite clasts than the gravel from Bone Cave and the lower gravel in the outer Reindeer Cave, yet a lower percentage of dolomite than the upper gravel unit. (N.b. the results of the lithological analysis of these gravels do not match those of Phemister (Lawson 1981b, and section 12.2.2.(b) above).)

There were therefore three different gravel units present in the Creag nan Uamh caves, separated on lithological grounds. This

Fig. 12.11 Lithological composition of gravel samples from Bone Cave and the inner chamber of Reindeer Cave.

(diagram overleaf)


D = dolomite
Q/T = quartzite/Torridonian
V = vosgesite
O = others
D
Q/T
V
O
BC3
BC7
BC9
BC10
RC11
%
0
20
40
60
80

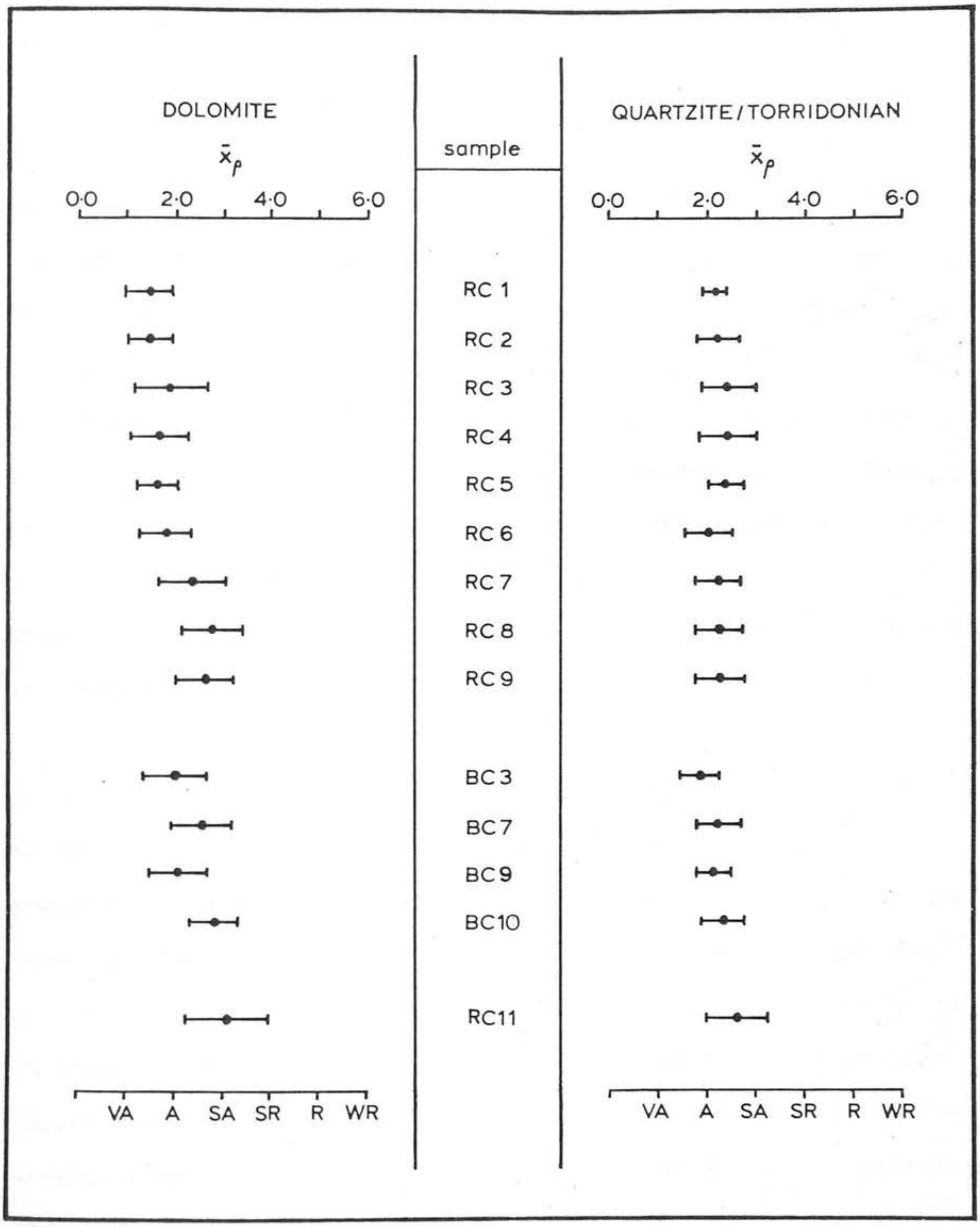
supports the view expressed by the previous excavators. The gravel in Bone Cave, which formed the lower gravel in Reindeer Cave, had the highest allochthonous content of the three gravel units. Of the other two units, the gravel occurring in the inner Reindeer Cave had the smaller autochthonous component. An explanation of this distinction between the three gravel units present must be sought in their mode of deposition and source of sediment supply.

It has been shown in chapter 10 that tills in the surrounding area possess very few dolomite clasts, and hence the presence of dolomite in the cave gravels is more likely to reflect an internal derivation. Results of the roundness analysis of both dolomite and quartzite/Torridonian stones in the -2ϕ to -3ϕ size fraction of certain of the gravel samples (Fig. 12.12) shows that the three gravel units also have different roundness characteristics. The dolomite clasts in samples RC7 to RC9 (i.e. the lower gravel unit), together with their correlatives in Bone Cave (BC3, BC7, BC9 and BC10), showed greater rounding than the upper part of the outer Reindeer Cave profile (samples RC1 to RC6). The sample from the inner Reindeer Cave (RC11) showed the greatest roundness of all the samples analysed. Therefore analysis of the autochthonous component of the gravels shows a clear difference between the gravels which existing sections indicate to have been water-lain (i.e. the lower gravel of the outer Reindeer Cave represented in Connecting Passage, and the gravel of the inner chamber) and the upper gravel unit which is composed predominantly of angular dolomite material. It is therefore suggested that this latter gravel unit was not fluvially deposited, but represents material that spalled from the roof and walls of the entrance chamber and was deposited subaerially. The results of the roundness analysis of quartzite/ Torridonian clasts are less clear (Fig. 12.12). Sample RC11 still

Fig. 12.12 Roundness analysis of dolomite and quartzite/Torridonian stones in gravel samples from Reindeer Cave and Bone Cave.

(N.b. the error bars represent $\pm$ 1σ from the mean.)

(diagram overleaf)


DOLOMITE
$\bar{x}_p$
0·0
2·0
4·0
6·0
sample
QUARTZITE / TORRIDONIAN
$\bar{x}_p$
0·0
2·0
4·0
6·0
RC 1
RC 2
RC 3
RC 4
RC 5
RC 6
RC 7
RC 8
RC 9
BC 3
BC 7
BC 9
BC 10
RC 11
VA
A
SA
SR
R
WR
VA
A
SA
SR
R
WR

showed the greatest degree of roundness, but the bipartite division of the outer Reindeer Cave section is not so clear and, if anything, showed a greater degree of rounding in the upper gravel. A Mann-Whitney U-test on these data showed no statistical difference at the 0.05 level between the roundness of quartzite/Torridonian stones of the upper and lower gravels. However, it must be stressed that the low number of quartzite/Torridonian particles counted from the upper gravel, necessitated by the low percentages of these lithologies present, renders these values statistically unrepresentative (Folk 1955). The presence of any allochthonous stones in the upper gravel, which ought to be purely autochthonous if it was a breakdown deposit, most likely represents a certain amount of mixing of the breakdown with the underlying fluvial gravel. This disturbance can be attributed to the presence of animals in the cave some time during or after the deposition of the upper gravel stratum, or to cryoturbation processes.

Differences in the lithological composition of the two fluvially-deposited gravel layers can be attributed to different transportation paths prior to deposition. The Bone Cave gravel was traced by the previous excavators from the entrance of the cave, through Connecting Passage to Reindeer Cave; the in situ section through this gravel in Connecting Passage attests to water flowing in this direction. In contrast, the higher percentage of dolomite stones and the greater degree of roundness of the clasts in the inner Reindeer Cave gravel suggest a longer transport path through a cave system prior to deposition. Imbrication of the pebbles, the position of the gravel close to the steeply sloping cave roof and wall and the limited spread of this deposit over the yellow silty-sands of the inner Reindeer Cave, suggest that it was deposited by a stream entering the chamber from below under hydrostatic pressure, and hence probably under phreatic conditions.

It is interesting to note that Stanton (1965) noted similar deposits in Gough's Cave, Cheddar, attributing a similar mode of deposition to them. The coarsening upwards is seen as a result of progressive deposition reducing the cross-section of the channel, effectively increasing velocity and thereby increasing the maximum size of particle transported. Such a stream therefore must have travelled through a cave system (believed to be one linking Uamh an Claonaite to the Creag nan Uamh caves: section 9.2.2), possibly for some distance.

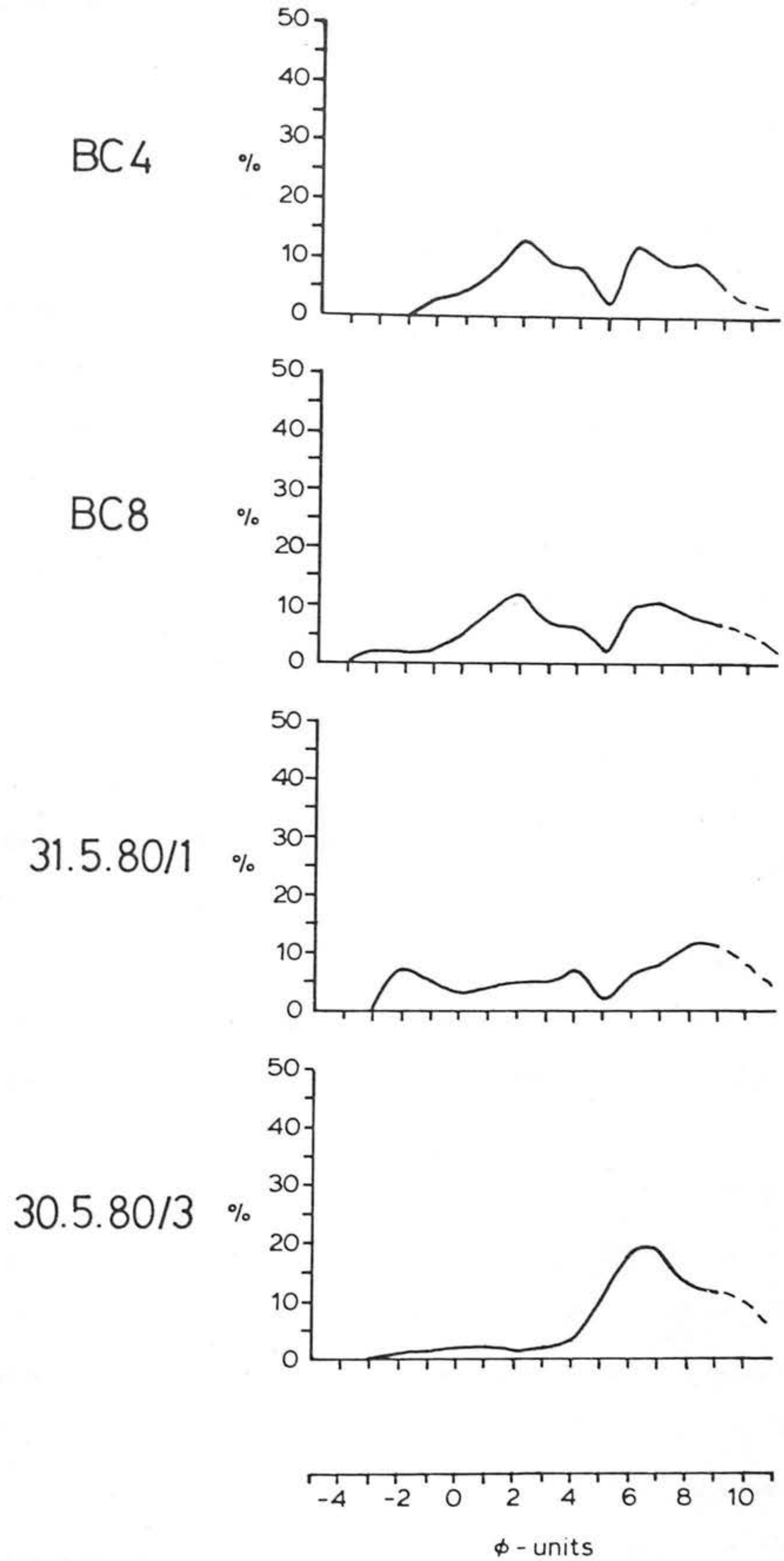
(c) The other deposits

Three other basic sediment types are present in the Creag nan Uamh caves: sticky muds, 'cave earth' and deposits referred to as 'fossil cave silts' in chapters 9 and 10.

Four samples of the sticky mud deposits in Bone Cave were analysed for their particle size distribution (samples BC4, BC8, 31.5.80/1 and 31.5.80/2). Samples BC4 and BC8 had similar particle size distributions (Fig. 12.13): they were essentially bimodal, with modes in the medium sands and medium silts (2ϕ and 6ϕ), and had platykurtic distributions (sensu Folk 1974). The other samples were somewhat different. Sample 31.5.80/1 was trimodal, with modes at -2ϕ, 4ϕ and 8.5ϕ, and had a platykurtic distribution; the -2ϕ mode relates to the numerous amphibian bones that this deposit contains. Sample 31.5.80/3, from pockets in the floor of Connecting Passage beneath the gravels, was leptokurtic and unimodal, comprising mainly medium silts (mode at 6.5ϕ). Little can be said about the mode of deposition as no sedimentary structures were seen in the present sections and none were reported by the previous excavators. It is suggested that they represent a 'wash' deposit, whereby groundwaters percolating through fissures in the bedrock wash fine particles from

Fig. 12.13 Particle size distributions of four samples of the basal 'sticky muds' (silty sands) from Bone Cave.

(diagram overleaf)


BC4
BC8
31.5.80/1
30.5.80/3
%
50
40
30
20
10
0
-4 -2 0 2 4 6 8 10
ϕ - units

soils or unconsolidated sediments above ground into the cave where they are deposited from roof-drips. The presence of two or more particle size modes in these sediments, plus their location in large entrance chambers, suggest that a secondary sediment population may be present, perhaps due to aeolian processes or the post-depositional translocation of fines from overlying deposits. Whatever their provenance or mode of deposition, these basal fine deposits are unrelated to the overlying gravel, as the latter occupied a discrete channel cut into them in Bone Cave (section 12.2.2) and there is a sharp contact between the silts and the gravel in Connecting Passage.

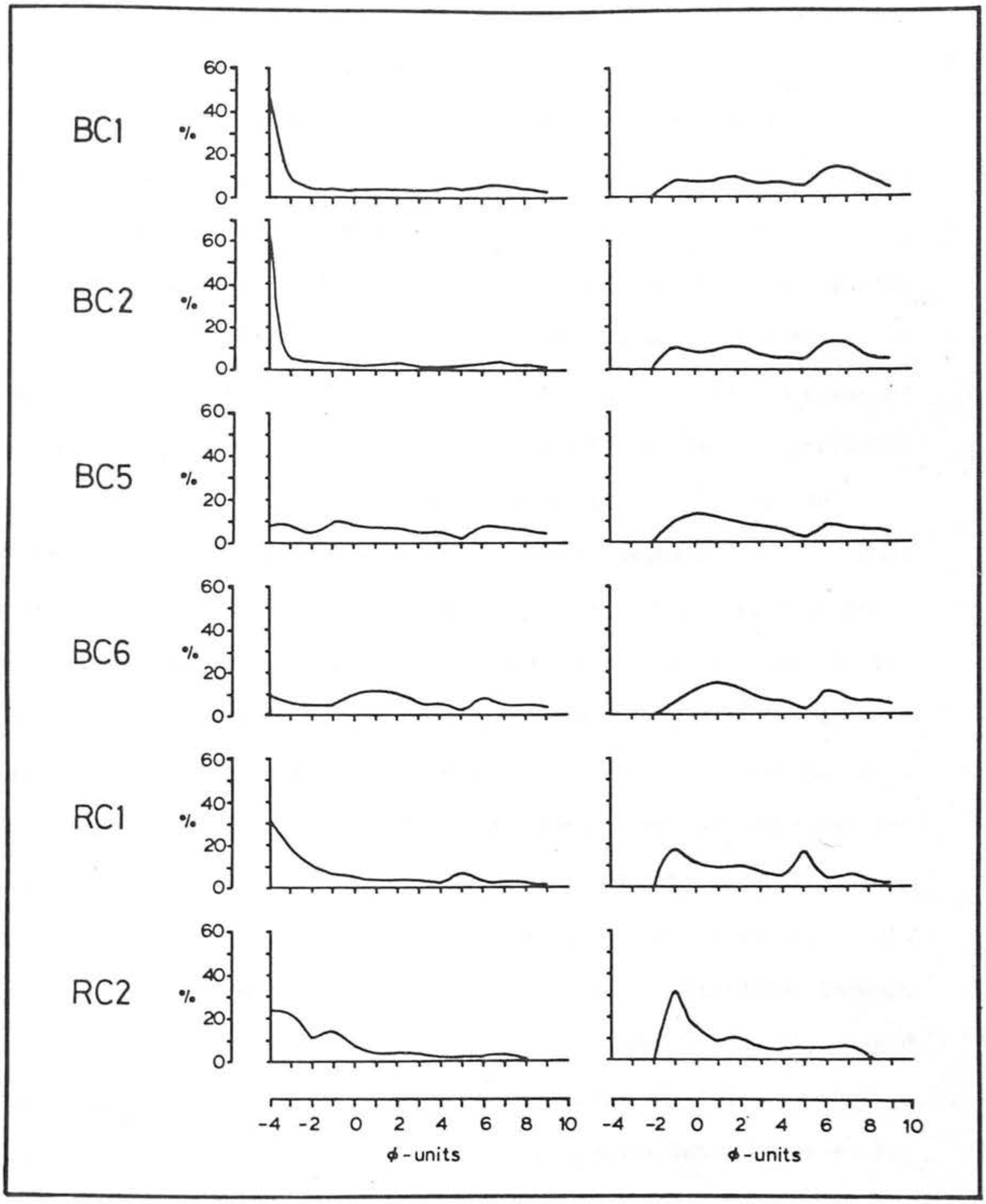
Since the earliest times of cave excavation, the literature has contained many terms to describe the mixed cave deposits which comprise varying quantities of breakdown fragments, mud and faunal remains, e.g. 'bone earth' (Pengelly 1864) and 'cave breccia' (Pengelly 1864; Dawkins 1874). 'Cave earth' is such a descriptive phrase, used in the present context by Peach & Horne (1892b, 1917) and retained by Callander et al. (1927); the term has been used in recent publications (e.g. Sutcliffe 1981), but its use should be discontinued, as it has no sedimentological meaning and no true 'earth' (in the sense of 'soil') can be said to exist in the cave environment where poor photic conditions prohibit the occurrence of most plant forms and severely restrict the faunal species present. In view of the varying nature of these types of deposit, the collective term 'cave entrance facies' is preferred (e.g. Kukla & Lozek 1958; Frank 1975), although retention of the expression 'cave earth' is permissible when no genetic significance is implied.

The 'cave earth' in the Creag nan Uamh caves is the most recent deposit present. It comprises many small dolomite splinters in a red or brown sandy-silt matrix and is rich in faunal remains. Four samples from these deposits in Bone Cave (samples BC1, BC2, BC5 and BC6) were analysed for particle size. It was assumed that the top two samples from the outer Reindeer Cave profile (RC1 and RC2) represented the 'cave earth' present in that cave. The results are presented in Fig. 12.14. Samples BC1 and BC2 proved to be the coarsest (median sizes of -3.4ϕ and -4.7ϕ respectively), the global particle size distribution showing one very large mode in the greater than -3ϕ size range. The particle size curves of the less than -2ϕ fraction show that there was a fairly even distribution of particle sizes, with small modes at -1ϕ, 2ϕ and 6.5ϕ. Samples BC5 and BC6 were from farther back in the cave, away from the extreme environmental conditions near the entrance where weathering processes would be more active. These samples were finer than the others (median sizes of 1.2ϕ and 1.4ϕ respectively), lacking the very marked coarse mode in their global particle size distributions. The Reindeer Cave samples fell between the other sample groups (median sizes of -2.9ϕ and -2.6ϕ respectively), being dominated by particles coarser than -1ϕ. These deposits are formed by a combination of chip breakdown (*sensu* Davies 1949) and the in-washing of fines by dripwaters as described above. The breakdown fragments account for the coarser modes, which is why the 'cave earth' could not be separated from the upper gravel layer on lithological grounds (section 12.3.2(a)).

The sediment source and mode of deposition of the fossil cave silts has largely been dealt with in chapter 10. Analysis of samples of these deposits from the Creag nan Uamh caves were included in that chapter and hence the main conclusions of section 10.4.1 apply here (*i.e.* the silts were derived from glacial abrasion products and deposited in flooded caves under an ice sheet). The presence of coarser deposits at discrete levels within the inner Reindeer Cave silts and the occurrence of more than one sedimentary unit suggest that the fine

Fig. 12.14 Particle size distributions of samples of 'cave earth' from Bone Cave and Reindeer Cave.

(diagram overleaf)



BC1
BC2
BC5
BC6
RC1
RC2
%
60
40
20
0
-4 -2 0 2 4 6 8 10
ϕ-units

sediments of this cave were not deposited in a single depositional phase. The slight colour difference between the pale yellow silts (10YR 6/3) and the underlying reddish-yellow silts (10YR 6/5) is not thought to signify two separate sedimentary units: the lower deposit possesses a greater range of particle sizes than the pale yellow silts, which may have affected the colour. However, the lowermost stratigraphic unit - the silty-clay and overlying manganese-stained gravel lens - is considered to be older. The slight red colouration of the silty-clay is unlike any of the deposits to which a glacial derivation has been applied, and the presence of manganese staining suggests the presence of peat above the cave as similar staining both of cave and surficial deposits is seen in the area at the present time. The complete decomposition of included breakdown fragments suggests that this unit is the oldest one so far discovered in the cave. The variation in particle size distributions, noted in the other caves as a whole, is more marked in these fossil silts from the Creag nan Uamh caves. Whereas similar deposits from other Assynt caves tend to be unimodal, the samples from the inner Reindeer Cave are multimodal and three from Badger Cave are bimodal. Possibly this may be due to wider fissures in the dolomite in the Creag nan Uamh area allowing a wider range of particle sizes to be washed into the flooded caves; filtering through smaller fissures would largely prohibit the incorporation of the sand modes present in some of the Creag nan Uamh cave samples. However, this locally increased width of sediment routeways into the caves is purely speculative.

12.3.3 Radiocarbon dating of the stratigraphy

In an attempt to obtain some definite dated horizons within the cave stratigraphy, osseous samples from two separate levels were sent to the N.E.R.C. Radiocarbon Laboratory at East Kilbride for radiocarbon assay. Sample RC/1 (n.b. not to be confused with the sediment sample of the similar sample number) was composed of fragments of reindeer antler from the upper gravel unit in the entrance chamber of Reindeer Cave. Sample RC/2 was a similar sample of reindeer antler fragments from the top 30 cm of the silts in the inner Reindeer Cave. Both samples were collected by J.E. Cree in 1926 and hitherto preserved in the Royal Scottish Museum. The samples had not been treated in any way, still being covered in a thin film of the cave sediments in which they originally lay. The protein fraction of the antlers was extracted for dating to avoid possible carbonate contamination due to burial in a carbonate-rich environment: the chemical pretreatment leaves no doubt that carbonate contamination was avoided (D.D. Harkness, pers. comm.). Sample RC/1 yielded a date of 10,080 $\pm$ 70 years B.P. (SRR-1788) and RC/2 a date of 18,040 $\pm$ 240 years B.P. (SRR-1789).

The RC/1 sample effectively dates the upper gravel layer of the outer chamber of Reindeer Cave to the end of the Loch Lomond Stadial. This fits in well with the postulated explanation of the upper gravel: frost-shattering of the walls and roof of the entrance chamber of the cave at this time would have resulted in such a breakdown deposit. This date also has significant implications for the early prehistory of Scotland (section 12.4).

The date obtained for sample RC/2 gives a minimum age for the silty deposits of the inner Reindeer cave and seems to support the idea of deposition during the existence of the last ice sheet. However, the date also has significant implications for the radiocarbon chronology of the Late Quaternary of northern Britain. Current opinion holds that the last ice sheet reached its maximal extent in Britain at, or slightly after, 18,000 years B.P. This is based on dates of 18,500 $\pm$ 400 (I-3372) and 18,240 $\pm$ 250 (Birm-108) from Dimlington, Humberside, a date of 18,000 $^{+1400}_{-1200}$ (Birm-146) from Cae Gwyn Cave, Tremeirchion, Clwyd, and dates of 18,900 $\pm$ 330 (Birm-213), 18,700 $\pm$ 500 (Birm-270a), 18,550 $\pm$ 185 (Birm-270b) and 18,400 $\pm$ 500 (Birm-270c) from Glen Ballyre, Isle of Man (Buckley & Willis 1969; Penny *et al*. 1969; Shotton *et al*. 1969; Shotton & Williams 1971, 1973). The Dimlington dates were from mosses located beneath Late Devensian till, and the Tremeirchion date from a mammoth bone sealed into Cae Gwyn Cave by till; these dates therefore suggest that the glacial maximum occurred later than *c*. 18,500-18,000 years B.P. The Glen Ballyre dates are problematic in that they were obtained from mosses overlying till of the last glaciation, suggesting ice advanced over this area prior to *c*. 18,900 years B.P., which directly conflicts with the date from N. Wales. Attempts to resolve this problem have tended to discount the Isle of Man dates as too old due to suspected hard water contamination (*e.g*. Shotton 1977).

The recent ice-sheet model of Boulton *et al*. (1977) typifies a concept that has become accepted by many workers in the British Quaternary, namely that the Late Devensian glacial maximum was synchronous throughout the whole of Britain. However, this is unlikely to have been the case if the recently proposed deglaciation chronology from deep-sea cores and palaeobiological evidence (summarised in section 6.4.5) is to be believed (Sissons 1981). Reindeer graze on lichens and mosses, and the RC/2 date (if accepted) indicates that reindeer were present in the Assynt area at *c*. 18,000 years B.P. This implies that large parts of the land surface in the area were ice-free by then, at a time when the last ice sheet was approaching its maximal extent farther south. It is consistent with a time-transgressive maximum ice advance, occurring first in the north of Britain and progressively later farther south.

12.3.4 Depositional history of the Creag nan Uamh caves

The Creag nan Uamh caves consist of large chambers that were formed phreatically some time long before the last glaciation, probably as part of a larger cave system since dissected by glacial erosion.

The earliest deposits visible today are silty clays containing breakdown material in the inner Reindeer Cave, over which extended at least one stream channel (Fig. 12.15(a)). The actual age and mode of accumulation of these deposits is unknown. The fine-grained deposits imply low energy conditions, and the manganese-staining of the gravels is similar to that which affects cave and surface streams elsewhere in the study area at present, suggesting an interglacial or at least interstadial age for these sediments. The sharp, irregular upper surface of the gravel layer reflects a break in sedimentation of unknown duration and subsequent erosion of these deposits.

Mixed sands, silts and clays of a reddish-yellow colour, containing some fallen roof-stones, were deposited on top of these sediments. A discrete layer of cobble-sized breakdown fragments separates this stratum from an overlying pale, structureless yellow silt which is devoid of breakdown material. In places where the marker-horizon of breakdown fragments is absent, the reddish-yellow silts merge imperceptibly into the pale yellow silts, which suggests that these sediments represent a single depositional phase (Fig. 12.15(c)). Sedimentation to the roof, to a level above that of the cave entrances, implies that these were blocked, presumably by ice, and the caverns totally flooded; deposition under an ice sheet is therefore invoked. A radiocarbon date of 18,040 $\pm$ 240 years B.P. (SRR-1789) provides a minimum age for these silty deposits.

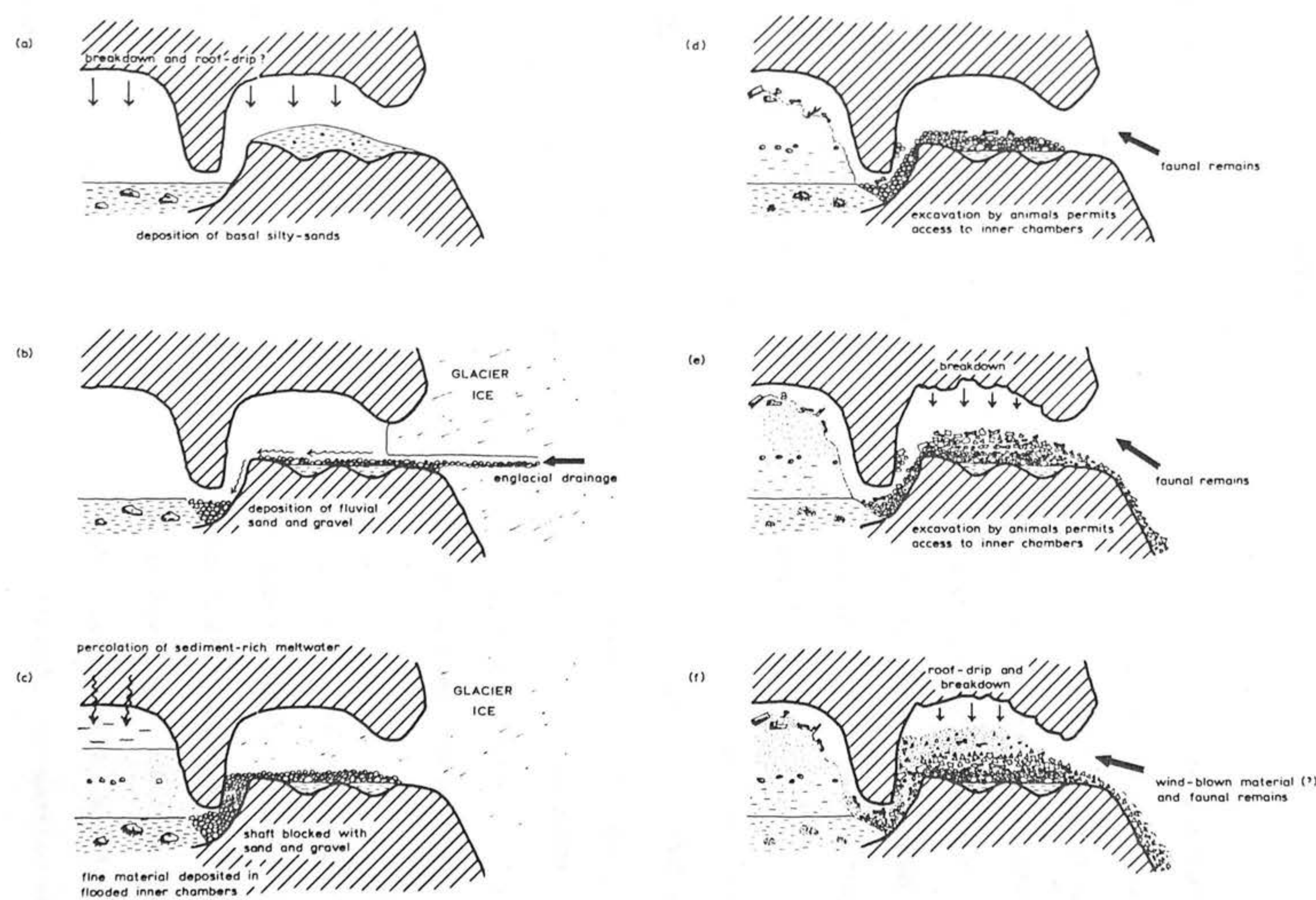
The last depositional phase in the inner Reindeer Cave was

Fig. 12.15 Schematic depositional history of the Creag nan Uamh caves.

(a) During the last interglacial, or during an interstadial during the last glaciation (?);
(b) during the build-up of the last ice sheet;
(c) during ice-sheet decay;
(d) deglaciation of the Allt nan Uamh valley completed;
(e) during the Loch Lomond Stadial;
(f) during the Postglacial period.

(N.b. the various arrows denote sediment inputs.)

(diagram overleaf)


(a)
breakdown and roof-drip ?
deposition of basal silty-sands
(b)
GLACIER
ICE
englacial drainage
deposition of fluvial
sand and gravel
(c)
percolation of sediment-rich meltwater
GLACIER
ICE
shaft blocked with
sand and gravel
fine material deposited in
flooded inner chambers
(d)
faunal remains
excavation by animals permits
access to inner chambers
(e)
breakdown
faunal remains
excavation by animals permits
access to inner chambers
(f)
roof-drip and
breakdown
wind-blown material (?)
and faunal remains

one of stream gravel, occupying a channel between the walls and roof of the chamber and the previously deposited silts. The gravel was introduced into the cave from below under hydrostatic pressure, and hence probably under phreatic conditions.

The age of the outer cave deposits in relation to those of the inner cave is difficult to determine. The crucial relationship is the one between the inner cave silts and the lower gravel unit of the outer Reindeer Cave. Little was noted by the original excavators about the junction between the two depositional units in the Reindeer Cave shaft. If the silts were the older deposit, one might expect the shaft and entrance chambers to have been filled with them, thus preventing the entry of the gravel found there in 1926. The filling of the shaft with gravel is therefore more easily explained by considering this phase to be earlier than the deposition of the silts (Fig. 12.15(b)). The unfossiliferous nature of the gravel is consistent with englacial meltwater drainage from ice occupying the Allt nan Uamh Valley in an early phase of the last ice sheet. The meltwater stream entered Bone Cave (Reindeer Cave possibly being blocked with ice at this time), flowed into Reindeer Cave by way of Connecting Passage, and away into the hillside.

The basal muddy silts of Bone Cave are older than these fluvial deposits and therefore probably pre-date the last ice sheet. They may relate to an interstadial during the last glaciation, and may be the same age as the basal fine deposits in the inner Reindeer Cave.

The upper gravel unit of Reindeer Cave represents extensive breakdown of the cave roof and walls (Fig. 12.15(e)). No mention was made as to whether the large amount of faunal remains found associated with this stratum was concentrated in any particular level, but the radiocarbon date of 10,080 $\pm$ 70 years B.P. (SRR-1788) probably gives a minimal date for this layer. The breakdown probably reflects intense frost-shattering in the entrance chamber of this cave during the Loch Lomond Stadial. At least towards the end of this period, the cave was occupied at different times by brown bears and possibly man (section 12.4). Evidence that might explain why similar deposits were apparently not represented in Bone Cave and Badger Cave has unfortunately been removed by the previous excavators.

The uppermost deposits in the entrance chambers of all the caves are entrance facies of varying amounts of small breakdown flakes in a red-brown silty matrix containing numerous bones - the so called 'cave earths' (Fig. 12.15(f)). These do not seem to be actively forming at present, but this may be due to recent disturbance within the caves. They are undoubtedly Postglacial in age.

12.4 ANALYSIS OF THE FAUNAL REMAINS

12.4.1 Introduction

The faunal remains referred to in this section are those collected during the 1926 - 1927 excavations, unless otherwise specified. It is hoped in future to be able to present a complete faunal list, but the necessary expertise was unavailable during the writing of this thesis.

There are three layers in the Creag nan Uamh caves which contain bones: the 'cave earth', the upper gravel unit and the top 30 cm of the silts in the inner chamber of Reindeer Cave.

12.4.2 Faunal assemblages from the Creag nan Uamh caves

(N.b. the remains listed below are class Mammalia unless otherwise specified.)

(a) Surface of fossil cave silts, inner Reindeer Cave

CARNIVORA	Ursus arctos	brown bear
	Alopex lagopus	arctic fox
	Felis lynx	northern lynx
	Canis lupus	wolf
ARTIODACTYLA	Rangifer tarandus	reindeer
RODENTIA	Dicrostonyx torquatus	arctic or collared lemming
	* Microtus cf. agrestis #	field vole?
	* Microtus cf. oeconomus	northern vole?
	* Arvicola terrestris	water vole
	* Apodemus sylvaticus	wood mouse

(b) Upper gravel unit, outer Reindeer Cave

CARNIVORA	Ursus arctos	brown bear
ARTIODACTYLA	Rangifer tarandus	reindeer
RODENTIA	¥ Dicrostonyx torquatus	arctic or collared lemming
	¥ Microtus gregalis	tundra vole
	¥ Microtus sp.	(a vole)

(c) 'Cave earth' in Badger Cave (1), Reindeer Cave (2) or Bone Cave (3)

INSECTIVORA	¥ Sorex araneus	common shrew (1)
	¥ a Soricid	(a shrew) (2)
PRIMATES	Homo sapiens sapiens	Man (1,2)
CARNIVORA	Felis sylvestris	wildcat (1)
	Mustela erminea	stoat (3)
	Meles meles	badger (1)
	Canis lupus	wolf (1)
	Vulpes vulpes	common fox (1,3)
	Ursus arctos	brown bear (1,2,3)
ARTIODACTYLA	Sus sp.	(a pig) (1)
	Cervus elephas	red deer (1,2)
	Capreolus capreolus	roe deer (1)
	Rangifer tarandus	reindeer (1,2,3)
	Ovis sp.	(a sheep) (1,2)
	Bos sp.	(an ox) (1,3)

RODENTIA	Arvicola terrestris	water vole (1)
	¥ Microtus cf. agrestis #	field vole (1)
	¥ Microtus oeconomus	northern vole (1)
	¥ Dicrostonyx torquatus	arctic or collared lemming (2)
LAGOMORPHA	Oryctolagus cuniculus	rabbit (1,3)
	Lepus sp.	(a hare) (1)

A variety of different unidentified bird species and fish, gastropods (including Cepaea nemoralis (2) and Patella sp. (2)) and amphibians (mainly frogs and toads) were also retrieved from this layer (c).

(Key: * identified by G.B. Corbet (British Museum, Nat. Hist.), October 1971; ¥ identified by A. Currant (British Museum, Nat. Hist.), August 1981, for the author; n.b. # Microtus cf. agrestis may be M. arvalis (common vole) (A. Currant, pers. comm.).)

A faunal list (identifications by E.T. Newton) was presented as an appendix in Peach & Horne (1917); these finds were from two layers in Bone Cave (layers 3 and 5 in Fig. 12.3). In view of the suspect nature of Peach and Horne's stratigraphy of this cave, it has been decided not to apportion species to a specific stratum, so those not already noted above are merely listed: weasel (Mustela nivalis), otter (Lutra lutra), rat vole (Microtus ratticeps), bank vole (Clethrionomys glareolus), chaffinch? (Fringilla coelebs?), barnacle goose (Branta leucopsis), mute swan? (Cygnus olor?), mallard? (Anas platyrhynchos), teal (Anas crecca), wigeon (Anas penelope), tufted duck (Aythya fuligula), long-tailed duck (Clangula hyemalis), eider duck (Somateria mollissima), common scoter (Melanitta fusca), ptarmigan (Lagopus mutus), red grouse (Lagopus lagopus scoticus), golden plover (Pluvialis apricaria), grey plover (Pluvialis squatarola), little auk (Alle alle), puffin (Fratercula arctica), frog (Rana temporaria), toad (Bufo vulgaris), natterjack toad (Bufo calamita), salmon or trout (Salmo sp.).

12.4.3 Palaeoenvironmental significance of the faunal assemblages

The faunal assemblage from the top of the inner Reindeer

Cave silts was dominantly composed of arctic or sub-arctic species. This accords with the radiocarbon date of 18,040 $\pm$ 240 years B.P. (SRR-1789) obtained from amongst this material, implying that the land surface around the caves was free of the glacier ice of the last ice sheet, although the climate would still have been cold and one cannot exclude the possibility of glacier ice existing elsewhere in the area. This assemblage could have continued to accumulate during the Windermere Interstadial until the inner cave was effectively sealed by deposition of the 'upper gravel' thermoclastic scree during the Loch Lomond Stadial; this would help to explain the presence of animals associated more with the 'taiga' and tundra belts today (e.g. arctic fox and wolf).

The upper gravel of the entrance chamber of Reindeer Cave yielded a similar faunal assemblage. Microtus gregalis is thought to have become extinct during the Late Glacial (e.g. Bramwell 1977; A. Currant, pers. comm.). The radiocarbon date of 10,080 $\pm$ 70 years B.P. (SRR-1788) from this layer relates to the end of the Loch Lomond Stadial and accords with the arctic fauna.

The 'cave earth' stratum contained many species that are still present in northern Scotland today. Other species no longer present probably owe their extinction to man rather than to the changing Postglacial climate (e.g. brown bear, wolf and reindeer). Much of the 'northern' character of the Postglacial assemblages can be explained by the geographical location of the Creag nan Uamh caves. Indeed, a region so far north in the British Isles is likely to have remained a refuge area for many arctic species trapped there as climate ameliorated. This may account for the presence of Dicrostonyx torquatus, and Microtus oeconomus is a noted survivor into the Postglacial, especially on damp ground (A. Currant, pers. comm.). The presence of

Cepaea nemoralis is interesting as this land snail is no longer found farther north than southern Skye (Kerney & Cameron 1979, distribution map 272). The numerous small frog bones in these deposits are due to the death of many of these amphibians whilst aestivating in the pockets of mud.

12.4.4 Archaeological significance of the Creag nan Uamh caves

Archaeological remains in these caves were scarce and nearly all restricted to the 'cave earth'. Diagnostic cultural remains were totally lacking (Lawson 1981b), yet the previous excavators in Reindeer Cave believed the upper gravel unit to have some archaeological significance. This stratum contained the remains of over 800 reindeer antler bases (*i.e.* representing over 400 individuals). The excavators claimed that most of the antlers had belonged to young animals and had been naturally shed, although several seemed to have been cut from the skull.

In order to assess the possible archaeological significance of this layer, it was decided to undertake an analysis of the reindeer remains. Sturdy (1975), using techniques developed by Bouchud (1966), was able to sex reindeer antlers by inspection and measurement. Similar techniques were employed in the present study, except that preservation of large sections of antler was poor and hence sexing was usually by inspection only. 474 antler bases were examined; all had been naturally shed except for 3 which appeared to have been deliberately cut. 36.5% proved to be male antlers, 55.9% were female, the rest being too worn to sex. The size of 13.3% of the antlers examined suggested that they had belonged to young animals. The antler bases were predominantly sub-cylindrical in cross-section, suggesting that the reindeer were of the migratory tundra-type (in contrast to the relatively sedentary woodland-type). It was evident that very few reindeer bones were present in the upper gravel layer and certainly not enough to account for the large number of antlers represented. It was also noted that very few sizeable pieces of the beam of the antler were present in the collection, which comprised mainly bases and tines that had been snapped from the beam. The beam of the antler was the portion preferred for the manufacture of implements such as javelin and harpoon points by Late Upper Palaeolithic man (Mellars 1974; Megaw & Simpson 1979; Morrison 1980).

The results of this analysis suggest that man was instrumental in the deposition of reindeer remains in the upper gravel layer of Reindeer Cave, collecting shed antlers from the surrounding area and leaving in the cave, during what was probably only a temporary occupation, those portions considered less useful for tool-making. Similar large amounts of reindeer antler fragments are known from caves elsewhere in Britain (e.g. Bosco's Den on the Gower, West Glamorgan (Falconer 1868, pp. 515-516), Banwell Bone Cave in the Mendips, Somerset (Sutcliffe 1955) and Tornewton Cave near Torquay, Devon (Sutcliffe & Zeuner 1962)).

If man's influence on the deposits of the upper gravel of Reindeer Cave is accepted, there are important implications for Scottish prehistory. Hitherto, the oldest archaeological sites in Scotland are the 'Obanian' shell middens of the west coast, dated at Lussa Wood, Jura, to c. 6,000 b.c. (Mercer 1974); a similar age was obtained from a charcoal sample associated with non-geometric microliths at Morton, Fife (Coles 1971). The date for sample RC/1 from Reindeer Cave therefore indicates occupation of the Creag nan Uamh caves some 2,000 radiocarbon years before the above Late Mesolithic sites. Although such a conclusion was considered unlikely on logistical grounds

(Lawson 1981b), the presence of Late Upper Palaeolithic man in the Assynt area around 10,000 years B.P. must now be considered a strong possibility.

12.5 CONCLUSIONS

The Creag nan Uamh caves have proved to be both palaeoenvironmentally and archaeologically important. The earliest deposits possibly date from the last interglacial, or at least from an interstadial during the last glaciation. Sediments relating to full glacial conditions, to the Lateglacial and to the Postglacial periods have been identified on the basis of geomorphic and sedimentological evidence, supplemented by two radiocarbon dates. A date of 18,040 $\pm$ 240 years B.P. (SRR-1789) from the top of the deposits in the inner Reindeer Cave suggests early deglaciation for the Assynt area and supports arguments for a time-transgressive glacial maximum. A date of 10,080 $\pm$ 70 years B.P. (SRR-1788) from a position thought to be near the top of a layer of thermoclastic debris supports a Loch Lomond Stadial age for this stratum. A large Lateglacial and Postglacial fauna, unique in Scotland, has been obtained from the later deposits in the caves. Certain evidence for the temporary occupation of Reindeer Cave by man <u>c</u>. 10,000 years ago, if accepted, makes this the oldest archaeological site in Scotland by some 2,000 years.

CHAPTER 13 : CONCLUSIONS AND SUGGESTIONS FOR FURTHER RESEARCH

13.1 THE ASSYNT AREA IN THE QUATERNARY PERIOD

Geomorphic features and sediments in the study area relating to periods pre-dating the last ice sheet are restricted to the major elements of the topography and to a few poorly-represented patches of red soil and cave deposits. Large-scale features such as corries have had a multiphase development and cannot be specifically related to a particular phase in the geomorphic evolution of the area. Elements of the landscape probably pre-date the Quaternary, but repeated inundation of the area by ice sheets in numerous glacial phases has resulted in the erosion and reshaping of such features. The dating of certain speleothems has shown that the main enterable cave passages in the area were in existence before the last glaciation. These caves offer the best opportunity for the discovery of deposits pre-dating the last ice sheet. $^{230}Th/^{234}U$ dates presented in this thesis (chapter 11) have shown that calcite speleothems were being deposited in certain Assynt caves during the last (Ipswichian) interglacial and during periods within the Early and Middle Devensian. Certain clastic cave deposits may also pre-date the last ice sheet.

A marked climatic deterioration heralding the onset of the last full glacial phase is well documented from a number of British sites. This climatic deterioration in the Assynt area seems to have occurred around 26,000 years B.P. At this time speleothem deposition ceased, suggesting the removal of a vegetation cover and the suppression of groundwater recharge (chapter 11). Palaeoclimatic inferences made from the distribution and size of former Lateglacial glaciers, which can be taken as a model of an early stage in the build-up of an ice sheet, suggest that the main snow-bearing winds came from the south and that glacial accumulation was greatly aided by winds from between south and west. Small glaciers developed first in the corries on the north and east side of the Assynt mountains and eventually coalesced. Continued ice build-up resulted in the overwhelming of the main north-south mountain ridge, and the formation of an ice sheet with an ice divide located somewhere to the east.

At the glacial maximum, ice flowed across the study area in an approximately east-west direction. Order-of-magnitude estimates suggest an ice surface at the ice divide of between 1,100 m and 2,100 m above present sea level. Therefore, the whole area, including the highest mountains, was buried by ice to a considerable depth. During the ice build-up the thermal regime of the ice was likely to have been temperate, but by the time of the glacial maximum that portion of the ice sheet over the study area was probably frozen to its bed. Clastic cave sediments have been interpreted as representing a period of increased cave-stream discharges followed by a phase when caves in the area were flooded, their entrances blocked by glacier ice. The earlier phase, during which coarse gravels containing large proportions of erratic lithologies were deposited, is believed to relate to an early stage of ice build-up. The later phase, when laminated, pale yellow fine sands and silts (the 'fossil cave silts', chapter 10) were deposited must relate to a time when basal meltwaters were present. Both types of clastic sediment were derived from local glacial deposits.

Modern views suggest that the last deglaciation was due to the starvation of the ice sheet rather than due to climatic warming. An early deglaciation of the Assynt area has been proposed, which started prior to 18,000 years ago if one accepts the radiocarbon date of 18,040 ± 240 years B.P. from fragments of reindeer antlers from Reindeer Cave (chapter 12). The deposition of certain coarse gravels in a number of Assynt caves seems to relate to the presence of large quantities of meltwater at this time. Palaeobiological analysis of the lake sediment stratigraphy of Cam Loch (e.g. Pennington 1977) suggests that deglaciation of the area was completed by 13,000 years B.P.

Vegetation of the area by ericaceous shrub species occurred in the Lateglacial Interstadial. Two $^{230}Th/^{234}U$ dates suggest that speleothem deposition was resumed in this period, but both dated specimens show high thorium levels which suggest the possibility of contamination, so such a conclusion is perhaps premature at present. During the ensuing Loch Lomond Stadial (c. 11,000 - 10,000 years B.P.), seven glaciers developed in the area (section 6.3). Outside the glacial limits there is evidence for intense periglacial activity. A layer of thermoclastic debris in Reindeer Cave appears to relate to this phase. It contains an arctic fauna and has yielded a radiocarbon date of 10,080 ± 70 years B.P. It is likely that man was present in the area, albeit on a temporary basis, at the end of this period.

The Postglacial climatic amelioration resulted in the rapid return of closed plant communities and the recommencement of speleothem deposition in the Assynt caves. Progressive vadose downcutting led to the establishment of the present subterranean drainage network. Afforestation of the area started c. 9,000 years B.P. with the development of birch woodland, subsequently overwhelmed by the development of blanket bog. Faunal assemblages show many 'northern' affinities, suggesting that the north of Scotland remained a 'refuge area' for many arctic and sub-arctic species well into the Postglacial period.

13.2 GENERAL CONCLUSIONS

A number of important points can be made at a more general level. It is appreciated that none of the following constitutes a major advance in the discipline and that these points are largely basic principles re-emphasised by this research.

(a) It has been shown that it is very necessary to consider a number of lines of evidence when attempting to elucidate the geomorphic evolution of an area. Certain topics of study support each other, helping to formulate broad conclusions (e.g. the study of glacial erratics and striae both suggest that former ice flow in the study area was from east to west); the same topics can also give important clues as to specific past conditions or events (e.g. the glacial striae give detailed information on former ice flow and signify temperate ice, and the erratics indicate an important early phase of W-E ice flow prior to the development of an ice divide).

(b) The influence of geological structure on landforms at a variety of scales has been emphasised on a number of occasions in this thesis, at the macroscale (e.g. the effect of zones of weakness in the bedrock on the orientation of glacial rock basins and ice-moulded bedrock protuberances), at the mesoscale (e.g. the effect of thrust faults on the development of many cave passages in the area) and at the microscale (e.g. the influence of structural bedrock weaknesses on the orientation of certain glacial friction crack forms).

(c) The Assynt area, like other parts of the country, bears witness to the marked effect that environmental conditions during the Loch Lomond Stadial have had on the distribution of some of the more impressive glacial and periglacial features of the landscape. The extreme climatic conditions at this time resulted in a period of geomorphic activity that, though short, was nevertheless intense.

(d) The importance of caves as traps for Quaternary sediments has been emphasised, and it is suggested that the Quaternary scientist should pay them more attention than has often hitherto occurred. Despite limitations, such as the complexity and fragmentary nature of cave stratigraphies, and the difficulty of relating cave sediment facies to other geomorphic evidence outside a cave, they can yield important palaeoenvironmental information.

(e) The use of $^{230}Th/^{234}U$ dating of speleothems for the first time in Scotland has highlighted the potential such a dating method has for extending radiometric chronologies back beyond the range of the ^{14}C technique. It is hoped that future advances will enable the large standard deviations associated with such dates to be reduced to a level more comparable with those of radiocarbon dates.

13.3 SUGGESTIONS FOR FUTURE RESEARCH IN THE ASSYNT AREA

No research topic of this nature can hope to cover adequately all aspects of the Quaternary geomorphology of a region.

The author believes that the major elements of the Quaternary geomorphic evolution of the Assynt district have been elucidated in this thesis, but suggests the following areas where further research would be beneficial.

(a) A thorough excavation of the inner Reindeer Cave would possibly yield further important palaeobiological information and achieve a greater understanding of the sequence and significance of the clastic sediments contained therein. Such an undertaking would require specialist equipment and supervision.

(b) The preliminary characterisation of certain glacial friction cracks (chapter 8) seems to suggest that these features warrant further detailed examination. Particular attention should be paid to the spatial distribution of friction cracks over glaciated rock surfaces, and especially those formed beneath Loch Lomond Advance glaciers: in these situations, it may be possible to relate size and shape characteristics to certain glaciological variables, such as ice thickness.

(c) Further dating of speleothem specimens would allow periods of no speleothem growth, and hence periods of continuous permafrost or probable glacial conditions, to be more accurately defined. There is an obvious need to look for samples that lie directly above and directly below the various sedimentary fills: the author has this problem partly in hand, for certain as yet undated specimens have already been collected. Isotopic analysis of speleothems may also eventually allow more detailed information of palaeotemperatures to be inferred from $^{18}O/^{16}O$ ratios of fluid inclusions in the calcite crystal structure. However, such analysis is at a very embryonic stage of development at the moment (T.C. Atkinson, pers. comm.).

(d) Despite a careful search, no evidence of a glacial limit comparable to the Wester Ross Readvance stage was found within the study area. It may be that glacier ice did not exist in the area at this time (or indeed that the area was totally ice-covered). Further work, particularly in the region to the south of the study area, may help to clarify this point.

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APPENDIX I

FURTHER RADIOMETRIC DATING OF MATERIAL FROM THE CREAG NAN UAMH CAVES

The significance of the date of 18,040 $\pm$ 240 years B.P. (SRR - 1789) obtained from a sample of reindeer antler fragments (sample RC/2) from the inner chamber of Reindeer Cave has been discussed in Chapter 12. Three further osseous samples from the same stratigraphic horizon in the cave were later submitted to the N.E.R.C. radiocarbon laboratory for dating in the hope that they would yield a similar age to sample RC/2, thus further supporting the argument for a time-transgressive Devensian glacial maximum. The age determinations were unfortunately not available when chapter 12 was typed, but are included here as they have some bearing on the age of the deposits in the caves and on certain comments made in section 12.3.3.

Samples RC/3 and RC/4 were single reindeer antler fragments and gave dates of 25,360 $^{+\,810}_{-\,740}$ (SRR - 2103) and 24,590 $^{+\,790}_{-\,720}$ (SRR - 2104) years B.P. respectively. Sample RC/5 was a small legbone of a juvenile reindeer, and yielded a date of 8,300 $\pm$ 90 years B.P. (SRR - 2105). The dates obtained from samples RC/3 and RC/4 lie close to the onset of glacial conditions in the Late Devensian glaciation of the area, as defined by the distribution of $^{230}Th/^{234}U$ dates from local speleothems (chapter 11). Sample RC/5 was much younger than the other dated material from the inner part of Reindeer Cave, and would seem to represent Postglacial material derived from the outer chamber by scavenging animals. The failure of all these samples to support the age determination obtained from sample RC/2 means that interpretation of the latter necessitates some caution. It must be stressed that sample RC/2 was composed of a number of antler fragments, and therefore the date of 18,040 $\pm$ 240 years B.P. may represent an 'average' date from a mixture of pre-glacial and post-glacial material, especially as material of widely differing ages seems to be present in this part of the cave.

Samples RC/2, RC/3, RC/4 and RC/5 all give minimum ages for the fossil cave silts in the inner Reindeer Cave, which now must be seen as having been deposited in water-filled caves under an ice sheet pre-dating that of the Late Devensian period. Other fossil cave silts in caves at lower levels in the area are still seen as being Late Devensian in age: these postulated age differences may have some bearing on the slight particle size differences between certain samples of these deposits from the Creag nan Uamh caves and samples from other caves in the study area (sections 10.3.1 and 12.3.2).

APPENDIX II

A LIST OF SEDIMENT SAMPLES ANALYSED DURING THIS RESEARCH

A total of 68 sediment samples were variously analysed for particle size distribution, lithological composition and roundness, not all of which are referred to in the text. This appendix lists all the analysed samples, in chronological order of collection (except for the Creag nan Uamh caves' samples which were collected in 1926-7 and are here listed after the samples collected by the author). The list is presented in two parts: firstly all the samples are listed, and then those samples whose size parameters are shown in the bivariate scattergram plots (Fig. 10.1).

(a) Analysed sediment samples

4.7.79/2a	Badger Cave: sand (10YR 7/6) from lowest exposure of deposits against W wall of outer chamber. Upper 7 cm of profile.
4.7.79/2b	Badger Cave: sand (10YR 7/6) from same exposure as sample 4.7.79/2a. Lower 7 cm of profile.
4.7.79/3	Badger Cave: silt and clay (7.5YR 5/6) from middle exposure of deposits against W wall of outer chamber. Lower 20 cm of profile.
4.7.79/4	Badger Cave: fine sand/coarse silt (10YR 7/6) from same exposure as sample 4.7.79/3. Upper 15 cm of profile.
4.7.79/5a	Badger Cave: silt (10YR 6/3) from highest exposure of deposits against W wall of outer chamber. Lower 9 cm of profile.

4.7.79/5b Badger Cave: silt (10YR 6/4) from same exposure as sample 4.7.79/5a. Upper 9 cm of profile.

6.8.79/1 Reindeer Cave: silt (10YR 6/3) from highest part of deposits in the inner chamber.

6.8.79/2 Reindeer Cave: fluvial gravel, composed of rounded stones, from section at S end of trench dug from foot of shaft into the inner chamber.

6.8.79/3 Reindeer Cave: inter-bedded fluvial fine gravel and sand overlain by rounded gravel (sample 6.8.79/2).

29.3.80/1 Traligill basin: silt (5Y 5/1) from base of solution features on part of the exposed Traligill Main Thrust plane surface (NC 268212).

18.5.80/1 Uamh an Claonàite: silty sand (10YR 4/4) from lower (NW) end of Mud Passage, above Sump 3.

18.5.80/2 Uamh an Claonaite: laminated sandy silt (10YR 5/4) overlain by silty sand (sample 18.5.80/1).

19.5.80/2 Uamh an Claonaite: fluvial gravel from active streamway on far side (NW) of Sump 1.

31.5.80/1 Bone Cave: sticky mud deposits (sandy silt) (7.5YR 5/5), containing small bones, from rear of the cave.

31.5.80/3 Bone Cave: silty clay from pockets in the bedrock floor of Connecting Passage, overlain by fluvial coarse gravel.

26.6.80/2 Lower Traligill Cave: angular dolomite-rich gravel in a fine matrix (7.5YR 5/4) overlain by a deposit of travertine, from niche to N of cave entrance, outside the cave.

14.8.80/1 Traligill basin: laminated, fine-grained alluvium (10YR 3/3) from a section in the bank of the Allt a' Bhealaich (NC 28101990).

14.8.80/2 Traligill basin: coarse till from section in bank of the Allt a' Bhealaich (NC 21852000).

15.8.80/1 Traligill basin: gritty sandy clay (10YR 4/4) from a section at NC 281198 in one of the flood channels N of Cuil Dubh Sink. Overlying dolomite bedrock and overlain by coarse diamicton and peat.

15.8.80/2 Traligill basin: sandy clay (10YR 4/2) overlying quartzite bedrock and overlain by peat at NC 27822071 in the Allt a' Bhealaich valley.

11.10.80/2 Reindeer Cave: pale yellow silt (10YR 6/3) from upper 45 cm of the deposits in the inner chamber, overlain by water-worn bedded gravel deposit (samples 6.8.79/2 and 6.8.79/3) and breakdown blocks.

11.10.80/3 Reindeer Cave: reddish-yellow silty sand (10YR 6/5) from 47-80 cm depth in the deposits exposed in the inner chamber.

12.10.80/1 Cnoc nan Uamh cave system: angular gravel in silty matrix (10YR 4/4) from Beechbarrow Chamber, Uamh an Tartair (Traligill).

12.10.80/2 Cnoc nan Uamh cave system: fluvial gravel overlain by silt in Landslip Chamber, Uamh an Tartair (Traligill).

12.10.80/3 Cnoc nan Uamh cave system: silt (10YR 6/4) overlying gravel (sample 12.10.80/2) in Landslip Chamber, Uamh an Tartair (Traligill).

11.4.81/1 Allt nan Uamh basin: laminated sand (10YR 3/2) from section in shakehole depression at NC 27122654 near Uamh an Claonaite.

11.4.81/2 Allt nan Uamh basin: sand (7.5YR 3/2) from same

	exposure as sample 11.4.81/1, overlain by the laminated sand.
11.4.81/4	Allt nan Uamh basin: silty clay (7.5YR 4/4) from section (flood resurgence) in shakehole depression at NC 27132653, near Uamh an Claonaite.
11.4.81/5	Allt nan Uamh basin: till overlying dolomite bedrock near sink for water from Loch an Claonaite (NC 27061617).
11.4.81/6	Allt nan Uamh basin: dolomite totally weathered to a silty clay (2.5YR 4/4), overlain by till (sample 11.4.81/5).
11.4.81/7	Allt nan Uamh basin: till from section exposed in bank of Allt nan Uamh at NC 26551723.
12.4.81/1	Traligill basin: till from slump depressions near Glenbain Cottage (NC 26412152).
12.4.81/2	Allt nan Uamh Stream Cave: silt (10YR 6/3) from ledges in Oxford Street, near Piccadilly Circus.
12.4.81/3	Allt nan Uamh Stream Cave: fine silt (10YR 5/4) from S end of Oxford Street.
28.5.81/1	Cnoc nan Uamh cave system: sandy silt (10YR 5/4) underlying stalagmite deposits on E side of the Stream Chamber, Uamh an Tartair (Traligill).
28.5.81/2	Cnoc nan Uamh cave system: laminated silt (10YR 5/4) at base of sediment sequence in dry passage on W side of the Stream Chamber, Uamh an Tartair (Traligill).
28.5.81/3	Cnoc nan Uamh cave system: gritty sand (10YR 4/4) unconformably overlying laminated silt (sample 28.5.81/2).
28.5.81/4	Cnoc nan Uamh cave system: gravel with sandy matrix (10YR 4/4) overlying gritty sand of sample 28.5.81/3.
28.5.81/5a	Cnoc nan Uamh cave system: sand (10YR 4/4), showing low

angle cross-bedding, overlying gravel that is probably the lateral equivalent of sample 28.5.81/4. Contains silt lenses.

28.5.81/5b Cnoc nan Uamh cave system: sandy silt (10YR 4/4) overlying sand of sample 28.5.81/5a.

28.5.81/6 Uamh Cailliche Peireag: mottled gritty silt (10YR 4/4 or 3/4) containing weathered dolomite.

29.5.81/1 Lower Traligill Cave: silty sand (10YR 4/4) recently deposited by flood water, N end of cave.

29.5.81/6 Tree Hole Cave: bedded fluvial gravel from bank cut by active streamway.

12.6.81/1 Allt nan Uamh basin: till from bank of small tributary of the Allt nan Uamh at NC 27771705.

14.6.81/1 Allt nan Uamh basin: till from section near fish farm, at NC 25341790.

11.7.81/1 Uamh an Claonaite: silt (10YR 5/4) from the Viaduct Series passages.

12.7.81/1 Cnoc nan Uamh cave system: laminated silt and clay (10YR 5/4) underlying gravel and sand (sample 28.5.81/4).

RC1 Reindeer Cave (1926): "1st 6" from top" (= 'cave earth', outer chamber) (7.5YR 3/2 moist)

RC2 Reindeer Cave (1926): "2nd 6" from top" (= 'cave earth', outer chamber) (5YR 4/6 moist).

RC3 Reindeer Cave (1926): "3rd 6" from top" (= 30-46 cm, outer chamber profile) (5YR4/6 moist).

RC4 Reindeer Cave (1926): "4th 6" from top" (= 46-61 cm, outer chamber profile) (5YR 4/3 moist).

RC5 Reindeer Cave (1926): "5th 6" from top" (= 61-76 cm, outer chamber profile) (5YR 4/6 moist).

RC6	Reindeer Cave (1926): "6th 6" from top" (= 76-91 cm, outer chamber profile) (7.5YR 5/4 moist).
RC7	Reindeer Cave (1926): "7th 6" from top" (= 91-107 cm, outer chamber profile) (10YR 5/3 moist).
RC8	Reindeer Cave (1926): "8th 6" from top" (= 107-122 cm, outer chamber profile) (10YR 5/4 moist).
RC9	Reindeer Cave (1926): "9th 6" from top" (= 122-137 cm, outer chamber profile) (10YR 5/4 moist).
RC10	Reindeer Cave (1926): gravel and sand from approx. 2 m depth in shaft at back of outer chamber.
RC11	Reindeer Cave (1926): rounded fluvial gravel, same as sample 6.8.79/2.
BC1	Bone Cave (1927): "1st 6" from top" (= 'cave earth') (7.5YR 5/6 moist).
BC2	Bone Cave (1927): "2nd 6" from top" (= 'cave earth') (7.5YR 5/5 moist).
BC3	Bone Cave (1927): "3rd 6" from top" (= fluvial gravel).
BC4	Bone Cave (1927): "4th 6" from top" (= basal clay ?).
BC5	Bone Cave (1927): 'cave earth' (collected by Abbé Breuil, "3 ft back on E side of cave").
BC6	Bone Cave (1927): 'cave earth' ("top layer, 20 ft back on W side of cave").
BC7	Bone Cave (1927): fluvial gravel ("2nd layer, 20 ft back on W side of cave").
BC8	Bone Cave (1927): silty sand beneath gravel ("3rd layer, 20 ft back on W side of cave").
BC9	Bone Cave (1927): fluvial gravel from near entrance to

Connecting Passage.

BC10 Bone Cave (1927): fluvial gravel from Connecting Passage.

(b) A list of the samples of fine-grained sediment, etc., whose size parameters were used to construct Fig. 10.1.

1. Fossil cave silt

4.7.79/4

11.10.80/2

11.10.80/3

12.10.80/3

12.4.81/2

12.4.81/3

28.5.81/2

30.5.81/2

11.7.81/1

12.7.81/1

2. Flood deposits in caves

18.5.80/1

18.5.80/2

29.5.81/1

3. Wash deposits in caves

BC4

BC8

31.5.80/1

28.5.80/6

4. Fine-grained cave stream deposits

28.5.81/1

28.5.81/3

28.5.81/5a

28.5.81/5b

cont./

5. Coarse fluvial deposits in caves	6. Fine-grained deposits on surface
6.8.79/2	29.3.80/1
6.8.79/3	14.8.80/1
26.5.80/1	15.8.80/1
12.10.80/1	15.8.80/2
12.10.80/2	11.4.81/1
28.5.81/4	11.4.81/2
	11.4.81/4
	11.4.81/6