LATE QUATERNARY GLACIATION AND ENVIRONMENTAL CHANGE IN SOUTHERN ROSS-SHIRE, SCOTLAND

C. Jill Tate

A Thesis Submitted for the Degree of PhD at the University of St Andrews

1996

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A thesis presented for the degree of Ph.D.
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C.Jill Tate
(M.A. St.Andrews)

1995
CONTAINS

PULLOUTS
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CHAPTER ONE

Introduction

1.1.
Despite considerable active research in the realm of Quaternary studies in Scotland, some parts of the Scottish Highlands remain largely uninvestigated in terms of both glacial and environmental history. For many areas of Scotland the glacial history has been examined: researchers have ascertained aspects of the nature of the Late Devensian (Weichselian) ice sheet that covered most of Scotland and subsequent readvances, chiefly the Loch Lomond Readvance. In a 1979 review of the Loch Lomond Readvance in the British Isles, Sissons published a map of the glacial limits relating to this latest period of glacial activity in Scotland (Figure 1.1).

Figure 1.1
The Loch Lomond Readvance limits in Scotland
(after Sissons, 1979e)
Since then this map has been only slightly modified with certain limits having been established in the western Grampians. For the area further north, little accurate information has been added to this picture. One of the most noticeable omissions in relation to the Loch Lomond Readvance is in the area between Glen Carron and Glen Shiel in southern Ross-shire. Published research regarding the earlier Late Devensian ice sheet is also sparse. Information regarding other aspects of the Quaternary such as those detailed in other parts of the Scottish Highlands are similarly lacking for this area. There are no published accounts of the vegetation history of the area nor of Late Quaternary sea-level changes. This study seeks to fill in some of these gaps with specific reference to the Late Devensian glaciation and aims to relate the patterns of glaciation with other contemporaneous environmental changes through to the establishing of interglacial conditions in the early Flandrian period. Recent studies have demonstrated that aspects of the Late Quaternary environment can usefully be related and this study attempts to follow a similar procedure.

1.2. Aims

The primary aim of this thesis is to establish the characteristics of Late Quaternary glaciation in southern Ross-shire and to ascertain the nature of contemporaneous environmental changes in the area. This is achieved through initial detailed geomorphological mapping of the area given in Figure 1.1. The principal aims in relation to the Late Devensian glaciation of the study area are to establish the limits of glaciers existing during the Loch Lomond Stadial and to reconstruct the surface morphology of the ice masses present. A secondary aim is to determine the patterns of ice movement and to make inferences regarding operative glaciological conditions from the mapped evidence in relation to both the Late Devensian ice sheet and the Loch Lomond Readvance. A third aspect of this research into past glacial conditions involves the calculation of former equilibrium line altitudes and glacier dynamics with a view to inferring palaeoclimatic conditions during the Late Devensian. The study also attempts to elucidate the nature of deglaciation of the Loch Lomond Readvance glaciers through geomorphological and associated biotic evidence.

The second major aim is to establish the nature of other major environmental changes relating to this period and to compare these with similar changes in adjacent areas. The first of these is concerned with vegetational development. Palynological investigations are undertaken to determine the nature of Late Devensian and early Flandrian vegetation development, to establish a chronology for the area by locating sites in stratigraphic positions adjacent to morphological features believed to maximal or recessional limits, in order to obtain a relative chronology for glacial stages and also to gain an insight into the nature and chronology of deglaciation in the early Postglacial period. The study also aims to compare biostratigraphic sequences in a regional context and to those from sites in Scotland.
The results of all three aspects of the Late Devensian environment are interrelated. A third aim is to elucidate the palaeoclimatic implications of the glacial, periglacial and biostratigraphic evidence. First to determine the relationship of the reconstructed glaciers to climate at their maximum extent and their response to climatic change, second to relate the periglacial evidence to climatic parameters in which they formed and third to derive climatic information from the nature of vegetation indicated by pollen analyses. These aims may be supplemented with a fourth - to place the evidence within a chronological framework.

1.3. Structure of thesis
The following chapter reviews previous literature in the North West Scottish Highlands in order to place the present study in context. Chapter three outlines the major characteristics of the study area and details previous work on the glaciation of the area and other and environmental changes. The methodology applied in the mapping of glacial and periglacial evidence is discussed in Chapter Four and the observed patterns of the most recent local glaciation are detailed in Chapter Five. Chapter Six describes and considers the implications of the evidence for the Late Devensian ice sheet.

This is followed by a chapter detailing the palynological investigations that were undertaken. Pollen analytical methods and concepts are first outlined, followed by the results of these analyses The evidence amassed in Chapters Four to Seven are discussed in relation to palaeoclimatic conditions in Chapter Eight and Chapter Nine attempts to bring together the various strands of geomorphological and biostratigraphic evidence to provide an overall assessment of the nature of the Quaternary environment in southern Ross-shire.
CHAPTER TWO

Late Quaternary glaciation and environmental change in the northern Highlands of Scotland.

2.1 Introduction.
The main aim of this chapter is to place the present study of glaciation and environmental change in southern Ross-shire in the context of work previously undertaken in Scotland. The review focuses primarily on research relating to the North West Highlands of Scotland but in some instances where a wider view is required, information from throughout Scotland is considered. In section 2.2, the chronological background in which glaciation and environmental changes occurred during the Late Devensian is discussed. This is followed by a main section on glaciation (2.3) then section 2.4, which considers other aspects of environmental change which occurred during the Late Devensian.

2.2. Chronology of Quaternary environmental change in Scotland.

2.2.1. Introduction.
Before the 1970s, Quaternary research in Scotland tended to focus on landforms and sediments associated with the last (Late Devensian) ice sheet and subsequent glacial readvances. More recently, however, there has been a steady increase in information regarding the pre-Late Devensian Quaternary in Britain. Investigations by the British Geological Survey over the last twenty years have considerably extended knowledge of the offshore stratigraphic record, and have also added more detail in terms of knowledge of the Late Devensian. The cumulative research both on land and offshore has enabled the establishment of a provisional chronology that extends to the Early Pleistocene although this is somewhat tentative in some respects, is incomplete and rests on information from only a limited number of sites and borehole records. An analysis of the global context in which Quaternary events in Britain occur provides a far more comprehensive chronology of Quaternary climatic variations, derived from oxygen isotope ratios of marine microfossils, associated sea-surface temperatures and from oxygen isotope ratios in ice sheets (Johnsen et al. 1972; Shackleton and Opdyke, 1973; Robin, 1983; Jouzel et al. 1987, 1990; Dansgaard et al. 1989; Oeschger and Langway, 1989; Alley et al. 1993). Figure 2.1 demonstrates the chronology of the Late Quaternary.
2.2.2. Pre-Late Devensian.

In general, evidence for Quaternary landforms and deposits that predate the Late Devensian is restricted to areas peripheral to the last (Late Devensian) ice sheet. Little is known of Early or Middle Pleistocene events in Scotland. Some information concerning pre-Late Devensian Quaternary stratigraphy has emerged from studies in terrestrial areas peripheral to the last ice sheet, but the terrestrial Early-Middle Pleistocene record is very fragmented and it has not been possible to date or correlate individual sites. A more complete stratigraphic picture is available from the evidence provided by the work of the British Geological Survey offshore in the North Sea Basin. This allows tentative correlations to be made with regard to establishing a terrestrial chronology. Sediments of demonstrably Quaternary age have been found to reach depths of up to 400 m in the North Sea Basin (Cameron et al., 1987), the lower part of which has been assigned to the Early Pleistocene and the upper part to the Middle Pleistocene, the two being
separated by the Matuyama-Bruhnes palaeomagnetic boundary (c. 730 ka BP). Events believed to be related to the Anglian Glacial (Cameron et al. 1987) and Wolstonian Glaciation (Stoker et al., 1985; Cameron et al., 1987) have been recorded and there is also evidence for an interglacial horizon that has been correlated with the Hoxnian Interglacial (Griffin, 1984; Cameron et al., 1987). Above this in stratigraphic sequences, borehole investigations have revealed evidence for Ipswichian Interglacial deposits.

The conventional boundary between the Middle and Late Pleistocene lies at the base of the last (Ipswichian) interglacial. Few sites in Scotland are attributed to this interglacial and none of these may be considered to be firmly established in terms of an Ipswichian Interglacial age. A number of sites indicate a possible Early-Middle Devensian glaciation (e.g. Stoker et al., 1985; Bowen et al., 1986; Sutherland, 1981; Connell and Hall, 1984). Nowhere, in Scotland, however, have glacial sediments been securely dated to the Early Devensian.

Although evidence from sites in England has enabled the identification of distinct interstadials during the Early and Middle Devensian, for example at Upton Warren (Morgan, 1973) and at Chelford (Worsley, 1977), in Scotland equivalent sites have not been unequivocally established. A limited number of sites with 'infinite' radiocarbon ages may be attributable to Early or Middle Devensian interstadials (cf. Sissons, 1981).

Within the global context, correlations have been made between the marine/terrestrial record and the oxygen isotope ratios obtained from polar ice sheets (Johnsen et al. 1972; Robin, 1983; Jouzel et al. 1987, 1990; Oeschger and Langway, 1989; Alley et al. 1993). This has yielded the possibility of refining the subdivision of the Quaternary although the correlation of the fragmented terrestrial sedimentary record in Scotland to the oxygen isotope record is still problematic, especially prior to the Late Devensian.

2.2.3. Late Devensian
The chronological framework for late Quaternary environmental change in Scotland outlined in the previous section identifies the former existence of several glacial periods. Several ice-sheet glaciations are recorded in other parts of Britain and elsewhere in the northern hemisphere, but in Scotland evidence for pre-Late Devensian glaciations is largely restricted to the North Sea Basin, N.E. Scotland and the Outer Hebrides, and the chronology of these glacial events is not firmly established. A more comprehensive and detailed chronology of glaciation is available for the Late Devensian and this is given in Figure 2.3. Climatostratigraphically, the Late Devensian in Great Britain has been conventionally divided into three chronozones (Gray and Lowe, 1977; Rose, 1985): the Dimlington Stadial of c. 26-13 ka BP, the subsequent Lateglacial or Windermere
Interstadial of c. 13-11 ka BP and finally the Loch Lomond Stadial dated at c. 11-10 ka BP.

**Figure 2.2**  
Location of Late Quaternary Sites in Scotland

The Dimlington Stadial was characterised by the growth and subsequent retreat of the last (Late Devensian) ice sheet. The following sections outline current information concerning the nature and chronology of first, Late Devensian ice-sheet glaciation, second, downwastage and associated readvances of the ice sheet, and finally, the widespread readvance of glacier ice that occurred during the Loch Lomond Stadial.

### 2.3. Late Devensian Glaciation in the Scottish Highlands and Islands.

#### 2.3.1. Introduction

The idea that a major ice-sheet glaciation of Scotland had been succeeded by an episode or episodes of more restricted glaciation was established well over a century ago (e.g. A. Geikie, 1863; J. Geikie, 1894). At the beginning of the present century, various memoirs of the Geological Survey devoted to the Northern Highlands (e.g. Peach et al., 1910,
1913 a, 1913 b) reaffirmed this viewpoint, and proposed a threefold sequence of glacial phases for the period now termed the Late Devensian (see section 3.3). Although it was later accepted that numerous pre-Late Devensian glaciations had affected Scotland, the main focus of subsequent research has been on the Late Devensian ice-sheet and its downwastage, and on the subsequent Loch Lomond Readvance, since only evidence relating to the Late Devensian period has been widely preserved throughout Scotland.

**Figure 2.3**

Chronology of Late Devensian Deglaciation
2.3.1. The Late Devensian Ice Sheet

Although the build up of the last Ice Sheet was traditionally associated with the beginning of the Late Devensian, c. 26 ka BP, recent research has introduced the possibility that glacier expansion may have been initiated during the Early Devensian, c. 75 ka BP (e.g. Sutherland, 1981, 1984; Bowen et al., 1986), although this remains to be firmly established. The last ice sheet in Britain is believed to have reached its maximum southern extent in England at c. 18 ka BP. At the stadial type site at Dimlington, east Yorkshire, (Rose, 1985; Wintle and Catt, 1985; Bowen et al., 1986) radiocarbon dates of 18,500 ± 400 yr BP and 18,240 ± 250 yr BP have been obtained from moss fragments included within silts underlying till (Penny et al., 1969) although it is not known how often nor to what extent glacier ice advanced and receded before reaching this maximum. In Scotland, Middle Devensian interstadial deposits at sites peripheral to the major highland areas indicate that coastal and lowland areas were ice-free between 35 and 26 ka BP. The chronology of the onset of glaciation is uncertain, but radiocarbon dated ice-transported mollusc fragments suggest that an independent Outer Hebridean ice-cap extended beyond the present coast after 23 ka BP (Sutherland and Walker, 1984). Sediments in the North Sea Basin (comprising part of the Marr Bank Formation) are known to have been deposited in a shallow-water glaciomarine environment to the east of the ice sheet margin (Sutherland, 1984a; Stoker et al 1985), a position considered to have been close to its maximum extent, radiocarbon dated at 17730 +/- 480 BP.

The culmination of the Late Devensian ice sheet has not been dated directly (see below) but stratigraphic evidence indicates that the expansion of ice nourished in the Southern Uplands occurred after the encroachment of Highland ice into the area, which suggests that the last ice sheet may have reached its maximum extent in north of Scotland somewhat earlier than in southern Scotland (Geikie, 1894; Sissons, 1967a, 1983; Sutherland, 1984) and the culmination of the ice sheet in northern Scotland may have occurred before (possibly long before) c. 18 ka BP. This pattern has been interpreted as reflecting a progressive southwards movement of the zone of maximum snowfall as the last ice sheet expanded, such that the ice sheet reached its maximum extent at its northern margins before it did so at its southern margins, an interpretation consistent with the southerly migration of the oceanic polar front during ice sheet build-up (Sissons, 1981).

The major ice shed of the last Scottish ice sheet extended from the north in Sutherland through the Rannoch basin to the Cowal peninsula in the SW. Highlands (Thorp, 1986). A second major ice mass developed in the Southern Uplands and subsidiary ice masses of sufficient magnitude to withstand the pressure of the main ice sheet occurred in Mull, Skye and Arran, the eastern Cairngorms, SE. Grampians and the Cheviots (cf. Sutherland, 1984). The movement of ice from these centres of accumulation has been
traced using erratics. In the Northern Highlands, for example, the westerly and north­westerly movement of the last ice sheet across Wester Ross and Assynt has been traced from the carry of the Moine rocks to the west of the Moine Thrust, of Torridon Sandstone across Lewisian Gneiss outcrops, and from the orientation of striae, which indicate the deflection of ice around some of the major mountains (Ballantyne et al., 1987). Although the precise position of the former ice shed over much of the Northern Highlands is unknown, a radial pattern of ice-flow towards the north and north-east near Caithness has been recorded (cf. Sutherland, 1984).

The lateral and vertical dimensions of the last Scottish ice sheet have been the subject of much debate and speculation, and are topics of recent and current research. They are considered in turn below.

2.3.2.1. Lateral extent of the last Ice Sheet

Until recently it was generally assumed that the Late Devensian ice sheet had covered the entire Scottish mainland. The last Scottish ice sheet was believed to have reached the edge of the continental shelf in the west, having moved westwards across the Outer Hebrides (Geikie, 1878; Jehu and Craig, 1923; Phemster, 1960), and to have extended eastwards and north-eastwards into the central and northern North Sea, where it was confluent with the Scandinavian ice sheet (Boulton et al., 1977; Andersen, 1981; Price, 1983). Sissons (1981) questioned the validity of this assumption but it was not until extensive stratigraphical investigations had subsequently been undertaken that his criticisms were partly substantiated (Cameron et al., 1987).

Although there is evidence that an earlier ice sheet extended as far west as St. Kilda (Sutherland et al., 1984; Bowen et al., 1986) and that in the North Sea basin a former ice sheet was once confluent with Scandinavian ice, recent seismo-stratigraphic research in the latter area together with stratigraphic information from NE Scotland has confirmed that the last Scottish ice sheet was much less extensive than previously believed. Parts of mainland Scotland and of the Outer Hebrides were considered to have remained ice-free during the Late Devensian (Sutherland, 1984; Bowen et al., 1986).

Sediments of Late Devensian age related to both the Scottish and Scandinavian ice sheets have, however, been found in the North Sea basin. A sequence of glacial deposits, consisting of an arcuate belt of moraine-like ridges and termed the Wee Bankie Formation, is situated 20 - 50 km east of the Fife and Angus coasts (Thompson and Eden, 1977; Figure 2.4). This has been interpreted as an end moraine marking the maximum eastward extension of the last Scottish ice sheet (Thompson and Eden, 1977; Holmes, 1977; Thompson, 1978). Glaciomarine sediments known as the Marr Bank
Formation occur outside this formation and are believed to have formed pro-glacially beyond the margin of the last ice sheet. Contemporaneous sediments of glacial and glaciomarine origin that are found farther north and east in the North Sea are believed to represent the Late Devensian (Late Weichselian) Scandinavian ice sheet (Cameron et al., 1987). At the Late Devensian ice sheet maximum, the southern and central North Sea Basin between Scotland and Scandinavia was apparently a shallow arctic marine embayment (Sutherland, 1984; Cameron et al., 1987), which may or may not have had a thick sea ice cover (Stoker and Long, 1984).

**Figure 2.4**

*Limits of the Late Devensian ice sheet in Scotland*

On land it was inferred that a large area of Buchan remained ice free throughout the Late Devensian. The stratigraphic sequence at Crossbrae Farm (Hall, 1984) appeared to
indicate that this area was not glaciated during the Late Devensian, confirming an earlier suggestion to this effect by Synge (1956, 1977). The extent of this ice-free area was not verified, but the wider stratigraphic evidence in Buchan strongly suggested that a substantial area was unglaciated and subject to severe periglacial conditions during this period (Hall, 1984; Sutherland, 1984) although this is inconsistent with the limit represented by the Wee Bankie moraine.

Similarly, the north-eastern corner of Caithness and the islands of Orkney was also considered to have escaped glaciation during the Late Devensian (Synge, 1977; Flinn, 1978). Recent evidence has suggested, however, that these areas were glaciated during this time and that the geomorphological evidence suggests a former area of cold based ice (Hall and Whittington, 1989; Hall and Bent, 1990; Brown, 1993). In Shetland there is a lack of dated deposits, but the radial patterns of erratic carry, roches moutonées and ice-worn grooves provides evidence for a local ice cap (Mykura and Phemister, 1976; Flinn, 1978). This ice cap apparently extended c. 60 km to the east of Shetland but was not in contact with Scandinavian ice, the intervening part of the North Sea basin being partly an arctic marine embayment and partly dry land (Long and Skinner, 1985; Cameron et al., 1986).

In the west, the Outer Hebrides are now believed to have supported an independent ice cap (von Weymarn, 1974, 1979; Coward, 1977; Flinn, 1978; Peacock, 1980, 1984), which was probably confluent with mainland ice, at least in the latitude of Skye and Wester Ross (Ballantyne, 1990). Stratigraphic evidence, amino acid ratios, and radiocarbon dates on marine transported shells have been used to establish that a small enclave in northern Lewis remained glacier-free during the Late Devensian (Sutherland and Walker, 1984). The ice sheet is now known to have terminated to the north of Lewis (Sutherland, 1991a) with a larger lobe extending towards the edge of the shelf to the south of the Hebrides (Selby, 1989; Peacock et al. 1992). Farther west, on St. Kilda, there was apparently only one or two small valley glaciers (Sutherland et al., 1984), which implies that ice from the Outer Hebrides terminated somewhere to the east of St. Kilda and that there may have been an area of dry, ice-free land on the western continental shelf (Sutherland et al 1984a; Sutherland 1984c, 1987d). It has been shown that the St. Kilda Basin was free of glacier ice by 15.25 ka BP. Seismo-stratigraphic investigations of offshore Quaternary deposits have, however, failed to reveal any unequivocal indications of the presence of absence of the Late Devensian ice sheet in this area (cf. Cameron et al., 1987). However, a recently identified periglacial trimline in the Trotternish Peninsula on Skye, inferred to represent the approximate upper limit of the last ice sheet, indicates that the margin of the ice sheet at its maximum extent lay 25 - 40 km to the north of Trotternish (Ballantyne, 1990), within a very short distance of the coast of Lewis.
2.3.2.2. Vertical extent of the Last Ice Sheet

It was traditionally assumed that the last ice sheet covered the entire Scottish mainland, including the highest mountains in the North West Highlands. This idea has persisted from the beginning of this century (e.g. Geikie, 1878; Peach et al., 1913a) and is still prevalent in some recent research. There have been various attempts to calculate the vertical dimensions of the last ice sheet using computer-based modelling, and three dimensional models of the former ice sheet have been produced. Such projections, however, have generally been based on the assumption of a laterally much more extensive ice sheet than is now known to have existed. Consequently, the calculations of Boulton et al., (1977, 1985) and Andersen (1981), for example, have resulted in an overestimation of the vertical dimensions of the ice sheet over Scotland during the Late Devensian although an alternative reconstruction by Boulton et al. (1985) may have more closely approximated its true extent. A theoretical profile of a former ice sheet of more restricted extent has also been calculated for an east-west transect across part of the North-West Highlands (Gordon, 1979). This model attempted to incorporate variables pertaining to the movement and nature of glacier ice and revealed some correspondence between areas where the ice sheet was predicted to have experienced basal melting and sliding and areas of scoured bedrock. This was only partially successful, however, probably as the model, although more realistic, overestimated the actual lateral limits of the ice sheet as determined by more recent research.

The more restricted dimensions of the lateral extent of the last ice sheet, together with the presence of ice free areas and coexisting ice caps such as that in the Hebrides implies an ice sheet of limited maximum thickness. This is consistent with models of glacio-isostatic rebound and associated observed patterns of raised shorelines (Lambeck, 1991a, 1991b).

Recent work involving the identification of periglacial trimlines in the North-West Highlands has produced limited, though significant evidence that indicates that the Last Scottish Ice Sheet achieved an altitude of between 700 - 800 m in northern Ross-shire (Ballantyne, 1984; Reed, 1988). Trimlines found on several hills in Torridon provide information as to the altitude of the ice sheet within this area and also give some indication of the maximum height that the ice sheet could possibly have attained (Ballantyne et al., 1987). Similarly, in Trotternish, Skye, a periglacial trimline, inferred to represent the approximate upper limit of the last ice sheet at its maximum thickness, declines in altitude northwards from 580 - 610 m to 440 - 470 m over a distance of 24 km. (Ballantyne, 1990). Basal shear stress values calculated from the inferred ice-sheet surface gradient and ice-sheet thickness fall within the range 30 - 39 kPa, much less than
values calculated for present day ice sheets. A possible interpretation of this apparent anomaly is that the Trotternish Peninsula was flanked by low-gradient ice streams flowing over deforming sediments (Ballantyne, 1990). This work would appear to suggest that ice-sheet reconstructions of the form of the last Scottish ice sheet that assume parabolic profiles may be invalid. This information appears to be consistent with the ice-sheet limits in Figure 2.4 and corresponds with the currently emerging view of an ice sheet of much reduced dimensions. Further south in the Ben Nevis area, however, evidence of striae and erratics indicate that the Late Devensian ice sheet exceeded 1110 m (Sissons, 1967; Thorp, 1987). These differences are considered further in Chapter Six.

2.3.3. Deglaciation and associated readvances of the Last Scottish Ice Sheet

Radiocarbon dating of organic deposits immediately overlying sediments laid down by the last Scottish ice sheet has provided some information regarding the timing of ice-sheet deglaciation in Scotland. A number of significant radiocarbon dates have revealed the point at which various localities in the Scottish Highlands became ice-free. Dates obtained on basal organic sediments at Callander (12,710 ± 270 yr BP; Lowe and Walker, 1974), Loch Etteridge (13,150 ± 390 yr BP; Sissons and Walker, 1974), Loch Droma (12,810 ± 155 yr BP; Kirk and Godwin, 1963) and at Cam Loch (12,956 ± 240 yr BP; Pennington, 1975) indicate that widespread deglaciation had taken place by 13-12 ka BP, particularly as the above-mentioned sites all lie within or on the fringes of the Scottish Highlands, fairly near to the sources of the last ice sheet. The radiocarbon dates suggest that the last Scottish ice sheet had largely wasted away by c. 12.5 ka BP.

The retreat of the last Ice Sheet was originally thought to have involved widespread stagnation and \textit{in situ} downwasting (e.g. Sissons, 1967a; Price, 1983). This long-held view has recently been challenged, particularly as this initial phases of ice-sheet retreat is now known to have occurred whilst Scotland was surrounded by polar waters (associated with the northward return of the polar front) until c. 13 ka BP (cf. Serjup \textit{et al}. 1987; Peacock, 1989; Peacock and Harkness, 1990). The climate was still extremely cold (McIntyre \textit{et al}., 1972) with temperatures falling to between -7°C and -9°C (Atkinson \textit{et al}. 1987). This is evident from marine sediments found principally along the east coast and in the North Sea Basin, which date from the period of deglaciation and which have been found to contain a high arctic micro- and macro-fauna. Similarly, although less convincingly, equivalent faunas have also been reported from the north-west of Scotland (Brady \textit{et al}., 1874; Gregory, 1980). Further evidence implying climatic severity during initial ice-sheet retreat is indicated by the occurrence of intra-formational fossil ice wedge casts in eastern Scotland (Anderson, 1940; Kirby, 1969; Greig, 1981; Gemmell and Ralston, 1984). Although much of the geomorphological evidence relating to the decline of the ice sheet may be interpreted as being indicative of stagnation (e.g. Price, 1983), it
may also be argued that the available landform evidence could equally be interpreted in terms of retreat of active ice, with stagnation only in areas where active ice lobes became cut off from their accumulation areas (Sutherland, 1984; Brown, 1983).

Active ice retreat is also supported by the discovery of evidence for glacial readvances that interrupted ice sheet deglaciation. Several such readvances have been reported. In the Northern Highlands, evidence for a readvance has been documented in Wester Ross. This involved the readvance of a lobe of ice nearly 25 km in width across Loch Gairloch, and, farther south across Loch Ewe and the surrounding low ground (Robinson and Ballantyne, 1979; Sissons and Dawson, 1981; Figure 2.5). The Wester Ross Readvance is believed to have occurred at c. 13.5 - 13 ka BP (Ballantyne et al., 1987) although is more likely to have been earlier. The status of the readvance is, however, uncertain, as little evidence for such an event has been discovered elsewhere.

**Figure 2.5**

*The Wester Ross Readvance (from Ballantyne et al., 1987)*

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O Depositional shoreline or shoreline eroded in drift  
■ Shoreline eroded in bedrock  
□ Upper limit of marine deposits  
□ Shingle ridge  
□ Other evidence for the marine limit  
18.0 Altitude in metres  
A See text

**AULTBEA**  
Moraine  
Limit of Wester Ross readvance  
Interpolated

Kilometres 0 20
The possibilities of other advances have also been considered. The glacial landforms and deposits at Achnasheen in the North West Highlands described by Peach et al. (1913b) in association with evidence for a glacially-dammed lake in Strath Bran, where a large tongue of ice overflowed from the Loch Fannich valley, were subsequently mapped by Sissons (1982b) and interpreted as readvance limits. Although Sissons initially suggested that this 'Achnasheen Readvance' correlated with the Wester Ross Readvance, it appears more feasible that the limits at Achnasheen were associated with a later readvance (Ballantyne et al., 1987) and was considered by Benn (1991) to relate to the Loch Lomond Readvance. Morphological evidence for two other postulated ice-sheet readvances, the Aberdeen-Lammermuir and Perth Readvances (Simpson, 1933; Sissons, 1963, 1964, 1967a) is no longer considered valid (e.g., Paterson, 1974; Sissons, 1974; Ballantyne and Gray, 1984).

Radiocarbon dating of organic deposits immediately overlying sediments laid down by the last Scottish ice sheet has provided some information regarding the timing of ice-sheet deglaciation in Scotland. A number of significant radiocarbon dates have revealed the point at which various localities in the Scottish Highlands became ice-free. At St. Fergus in Aberdeenshire, glaciomarine sediments, deposited during retreat, have yielded a radiocarbon date of 15320+/− 200 BP (Hall and Jarvis, 1989). Dates obtained on basal organic sediments at Callander (12,710 ± 270 yr BP; Lowe and Walker, 1974), Loch Etteridge (13,150 ± 390 yr BP; Sissons and Walker, 1974), Loch Droma (12,810 ± 155 yr BP; Kirk and Godwin, 1963) and at Cam Loch (12,956 ± 240 yr BP; Pennington, 1975) indicate that widespread deglaciation had taken place by 13-12 ka BP, particularly as the above-mentioned sites all lie within or on the fringes of the Scottish Highlands, fairly near to the sources of the last ice sheet.

The radiocarbon dates described above suggest that the last Scottish ice sheet had largely wasted away by c. 12.5 ka BP. Although there has been considerable debate as to whether or not glacier ice survived in mountain areas throughout the Lateglacial Interstadial (e.g., Sissons, 1967a, 1979b; Sugden, 1970) there is no conclusive evidence to support either argument. The magnitude of climatic amelioration (Coope, 1977a, 1977b; Coope et al., 1977) in England and southern Scotland however, renders it improbable that glaciers survived throughout the Lateglacial Interstadial, except possibly in high corries in the Scottish Highlands (Ballantyne et al., 1987; Sutherland, 1991).

It is not known whether complete deglaciation of Scotland occurred during the Lateglacial Interstadial as the evidence based on environmental conditions is equivocal (cf. Sutherland, 1984; Bowen et al., 1986). Sugden (1970) envisaged that ice may have persisted throughout the Lateglacial Interstadial in the Cairngorms, but the limited extent
of Loch Lomond Readvance glaciers in many areas suggests that they formed *ab initio* during the Loch Lomond Stadial. Conversely, the concept of complete deglaciation during the interstadial was supported by Sissons (1974, 1976) who argued that most of Scotland was free of glacier ice by 12.5 ka BP. However the considerable masses of glacier ice that accumulated during the Loch Lomond Stadial in parts of the western and central Highlands have suggested to some writers that glacier ice may have survived the Lateglacial Interstadial (Peacock, 1970; Sugden, 1970) and recent work by Sutherland (1987) advocates the survival of glacier ice in high corries in Wester Ross.

The rapid climatic amelioration with which the beginning of the Lateglacial Interstadial is associated apparently coincided with the northwards retreat of polar waters from the Atlantic Ocean, which occurred at c. 13.5 ka BP (Ruddiman and McIntyre, 1973; 1981a; Ruddiman et al. 1977) with North Atlantic Drift waters reaching the Scottish coast (Peacock and Harkness, 1990). This change in climatic conditions has been identified on land both from palynological evidence and from the environmental implications of contemporaneous coleoptera assemblages although this data is from southern Scotland and England.

Evidence from pollen studies indicates that the onset of the Lateglacial Interstadial was characterised by the establishment of a pioneer vegetation cover. Radiocarbon dates from a number of sites in Scotland have indicated that this initial colonisation occurred at c. 13.1 - 12.8 ka BP (Bishop, 1963; Kirk and Godwin, 1963; Sissons and Walker, 1974; Pennington, 1975; Walker and Lowe, 1982). Following this pioneer stage various regional patterns are known to have developed. These comprised mainly grassland and dwarf shrub heath, some birch and variable amounts of *Empetrum, Juniperus* and *Salix*. The sequence and chronology of Lateglacial Interstadial vegetation is complex and is considered in more detail in section 2.5.

Studies of Lateglacial Interstadial coleoptera have also revealed information regarding climatic change during this period. Coleoptera provide a sensitive indicator of climatic fluctuations and respond quickly to environmental change (Coope, 1977a, 1977b; Coope et al., 1977; Atkinson et al., 1987). Bishop and Coope (1977) were able to detect a rapid influx of thermophilous coleoptera in early Lateglacial Interstadial deposits at sites in SW. Scotland. This, they suggested, was indicative of a rapid climatic amelioration with mean July temperatures reaching c. 15°C (at least as warm as the present day). Coleoptera assemblages higher in the interstadial deposits indicate a subsequent decline in mean July temperatures, at or slightly before c. 12 ka BP, to c. 12°C. This was thought to have been followed by a slower deterioration of summer temperatures between 12 - 11 ka BP (Bishop and Coope, 1977). Only a limited amount of research has been
undertaken on Lateglacial Interstadial coleoptera in Scotland, however, and the chronology of climatic change during this period is primarily based on palynological evidence. Water temperatures at c. 12.8 ka BP were calculated to be within 1 - 2°C of those of the present, from analysis of marine faunal assemblages (Peacock, 1981b, 1983a; Peacock and Harkness, 1990).

There have been a number of attempts to subdivide the chronology of the Lateglacial Interstadial (e.g. Pennington, 1975) but the lack of correlation between chronostratigraphic and litho-, bio-, and climato-stratigraphic boundaries has prevented the widespread adoption of proposed subdivisions. Some evidence has been found, however, for a brief climatic reversal in Scotland during the interstadial between c. 12 ka BP and c. 11 ka BP. This climatic oscillation has been equated with the Older Dryas chronozone identified in Northern Europe (c.f. Lowe and Walker, 1977 and section 2.5) although its status in Scotland is controversial. Whilst some palynological studies have indicated that a unidirectional vegetation succession occurred without interruption during the Lateglacial Interstadial in Scotland from c. 13 ka BP until the beginning of the Loch Lomond Stadial (Walker and Lowe, 1977) increasing evidence for a minor climatic recession has been found at sites in Scotland (Donner, 1957; Vasari and Vasari, 1968; Clapperton et al., 1975; Pennington, 1975; Benn et al 1992). There is, however, no indication of a climatic reversal at this time in the evidence from available coleoptera studies (Coope, 1970, 1977a, 1977b, Bishop and Coope, 1977) and it may be adduced that any climatic oscillation that may have occurred was of short duration, of low amplitude and may have been only local in effect (Walker and Lowe, 1990). It is now possible, however, to place these changes in the context of climatic changes in the northern Hemisphere and to attempt to correlate events in Scotland with those recorded in the Greenland ice cores (e.g. Alley et al 1993) which show some evidence of cooling during the Older Dryas and confirm the importance of thresholds or triggers in the North Atlantic climate system to changes in regional climate.

A decline in temperature (between c. 11.5 - 11 ka BP) prior to the Loch Lomond Stadial is indicated by both vegetational and coleoptera evidence.

The end of the Lateglacial Interstadial is not firmly dated, but the boundary between this chronozone and the later Loch Lomond Stadial is generally accepted as c. 11 ka BP (Gray and Lowe, 1977).

2.3.4. Loch Lomond Readvance.
A sharp change in environmental conditions marked the beginning of the Loch Lomond Stadial, the last stage of the Late Devensian. A climatic deterioration occurred, probably
as a result of the renewed southward migration of the oceanic Polar Front (Ruddiman and McIntyre, 1981). This was accompanied by a readvance of glacier ice, a marked vegetational revertance to an open vegetation cover, increased slope instability (Bowen et al., 1986) and greater periglacial activity (Ballantyne, 1984).

The precise timing of the Loch Lomond Stadial is uncertain, as evidence for the timing of the various aspects of environmental change is only broadly coincidental. The change in climate at c. 11 ka BP was reflected in a marked change from thermophilous to arctic coleoptera assemblages in SW. Scotland (Coope, 1977b). Evidence of this nature indicates that average July sea-level temperatures fell from c. 12°C at c. 11 ka BP to less than 9°C by c. 10 ka BP before rising rapidly to c. 15°C by c. 9 ka BP (Bishop and Coope, 1977). The deterioration in climate has been clearly identified in the pollen stratigraphic record (cf. Walker, 1984). Radiocarbon dates from a number of palynological sites imply that severe climatic conditions had been established well before 10.7 ka BP (e.g. Vasari, 1977; Walker and Lowe, 1982). This is described in greater detail below (Chapter Seven).

The period of renewed glaciation termed the Loch Lomond Readvance (Simpson, 1933) was assigned to Godwin’s pollen zone III (c. 10,750 - 10,250 yr BP; Donner (1957). In a wider context this period is considered equivalent to the Younger Dryas, chronozone of Northern Europe. Donner carried out palynological investigations both inside and outside associated glacial limits. The timing of glacial expansion is, however, not confidently established. A number of dates are available from marine clays containing shells that have been transported or overridden by Loch Lomond Readvance glaciers. Radiocarbon dates from eight sites have ages ranging from 12.3 ka BP to 10.9 ka BP (Sissons, 1967b; Gray and Brooks, 1972; Peacock, 1971c; Rose, 1980; Browne, and Graham, 1981). These dates provide a maximum age for the expansion of Loch Lomond Readvance glaciers (with most dates falling within the period 11.8 - 11.3 ka BP). This indicates that the culmination of the readvance occurred after 10.9 ka BP (Sutherland, 1984). The available dates indicate a glacial maximum between 10.9 and 10.7 ka BP (e.g. Lowe and Walker, 1976; Lowe, 1978; Walker and Lowe, 1979). The synchronicity of the readvance maximum in different localities has not yet been established (Bowen et al., 1986). However, a direct date for the culmination of the Loch Lomond Readvance has recently been reported by Rose et al. (1988). This evidence indicates that the Loch Lomond glacier reached its maximum extension sometime after 10.5 ka BP, a date conflicting with the view that the Loch Lomond Readvance glaciers associated with the western Grampian ice cap had wasted from their source areas by this time (e.g. Lowe, 1987; Lowe and Walker, 1980, 1981, 1984; Tipping, 1985; Dawson et al., 1987) and suggests that the
maximum expansion of glaciers was relatively late in the Loch Lomond Stadial (Rose et al., 1988).

Although it has not proved possible to date the end of the Loch Lomond Stadial accurately, but the combination of radiocarbon dates and palynological investigations has indicated that following this period, a rapid climatic amelioration occurred, which marked the beginning of the Flandrian. The earliest Flandrian dates indicative of ice retreat and the culmination of the Loch Lomond Stadial at various sites range between 11.6 ka BP and 8 ka BP (e.g. Sissons and Walker, 1974; Pennington, 1977a; Vasari, 1977; Lowe and Walker, 1977, 1980). However most of the available dates fall between 10.9 ka BP to 9.3 ka BP and the end of the Loch Lomond Stadial has been assigned to this period (Price, 1983).

The occurrence of a widespread advance or readvance of glaciers in Scotland after the general retreat and downwastage of the last ice sheet has been recognised for nearly a century and a half (e.g. Forbes, 1846; Chambers, 1855; Maclaren, 1855; Geikie, 1863), and was first given the name 'Loch Lomond Readvance' by Simpson (1933). During the past three decades a great deal of research has been devoted to establishing the extent and chronology of this readvance, and to modelling the climate of the associated stade (the Loch Lomond Stadial) on the basis of reconstructions of contemporaneous glaciers.

Despite recent research, the extent of the Loch Lomond Readvance remains undetermined for considerable areas of the north-west Highlands and is elsewhere uncertain; the precise timing of the initiation and culmination of the readvance and associated stadial are still controversial and the nature of the contemporaneous climatic regime has not been established with certainty.

2.3.4.1. Evidence.
Various forms of evidence have enabled the detailed reconstruction of Loch Lomond Readvance glaciers. With regard to delimiting the lateral extent of former glaciers, end moraines have been the most extensively-used form of evidence. They are most prevalent where they mark the limits to small Loch Lomond Readvance corrie glaciers (e.g. Sissons, 1977a). These landforms vary markedly in size and form, ranging from minor ridges of boulders to massive accumulations of boulder-studded, loose-textured drift. Other glacial landforms used to determine the lateral extent of readvance glaciers include lateral moraines, drift limits, limits to hummocky drift and limits of glacially-transported boulders (e.g. Sissons, 1977a; Ballantyne, 1989). Certain ice-contact fluvioglacial forms have also been used to this end (e.g. Sissons, 1977b) in addition to downslope limits of some types of periglacial features, especially boulder lobes (Sissons and Grant, 1972;
Sissons, 1977b). More, recently the use of periglacial trimlines has proved to be a successful method of defining the upper limits of former Loch Lomond Readvance glaciers (Thorp, 1981, 1986; Ballantyne, 1989) and Thorp (1981) and Ballantyne (1982) have contrasted the degree of weathering inside and outside readvance limits on spurs and rock surfaces. The concept of 'trimlines' is considered further in Chapter 4. Striae, erratics, fluted and streamlined moraines, ice-moulded bedrock and roches moutonées indicate the direction in which Loch Lomond Readvance glacier ice moved (e.g. Peacock, 1967; Sissons, 1967a, 1977a). A change in the direction of these features may also define the position of a former glacier limit.

Where glaciers of Loch Lomond Readvance age terminated in the sea, their limits may be recorded by a drop in the marine limit (the highest limit of marine action) (e.g. Ballantyne, 1989). Stratigraphic and biostratigraphic evidence have also been employed to establish former glacier limits. This includes sedimentological evidence based on the comparison of stratigraphies within and outside glacial limits: Lateglacial Interstadial organic sediments are found only at sites outside the limits of the Loch Lomond Readvance (e.g. Robinson, 1977; Walker et al., 1988).

2.3.4.2. Distribution.
Using the criteria described above, the limits of the Loch Lomond Readvance have been mapped over broad areas of Scotland. The results of this mapping have provided a fairly comprehensive reconstruction of former glaciers in the western Grampians (Peacock, 1970b; Thompson, 1972; Sissons, 1979; Thorp, 1981, 1986), some parts of the southern and eastern Grampians (e.g. Sissons, 1972, 1974; Sissons et al., 1973; Sissons and Sutherland, 1976) and the Hebrides (Gray and Brooks, 1972; Ballantyne and Wain-Hobson, 1980; Ballantyne, 1989). The extent of Loch Lomond Readvance glaciation has also been determined for parts of the Northern Highlands (Sissons, 1977a; Robinson, 1977; Ballantyne 1986a; Lawson, 1986; Reed, 1988). Figure 2.6. shows the distribution of Loch Lomond Glaciers in the western Scottish Highlands as far as it is known. The distribution and altitude of reconstructed of Loch Lomond Readvance glaciers have been used in regional palaeoclimatic reconstructions; this topic is considered in section 2.6.

It has been established that a major icefield developed as a centre of ice accumulation in the western Grampians (Thorp, 1986) and fed major outlet glaciers occupying the Rannoch, Ossian, Erich and Treig valleys. Thorp (1986) has demonstrated that the maximum thickness of ice over Rannoch Moor was in excess of 600 m. Ice emanating from high ground to the west of the Great Glen gave rise to a number of valley glaciers that extended along Glens Arkaig, Lochy and Garry to occupy the western part of the Great Glen (Sissons, 1979b). Further centres of ice dispersal existed between Glen
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The existence of outlying icefields, corrie and valley glaciers has been demonstrated in other areas of Scotland. In the Cairngorms and south-east Grampians, the extent of the Loch Lomond Readvance has been the subject of much debate (Sugden, 1970, 1971, 1973, 1977; Sissons, 1967, 1972, 1973, 1979a) centred on the nature and significance of hummocky moraine for delimiting Loch Lomond Readvance glaciers. It was concluded that this area supported only small corrie and valley glaciers during this period. Evidence for Loch Lomond Readvance glaciers has been found in the Inner Hebrides, on Arran, Rhum, Mull and Skye. A small ice cap existed in eastern Mull, to the west of which were four small independent glaciers (Gray and Brooks, 1972). Several small valley and corrie glaciers developed on Rhum (Ballantyne and Wain-Hobson, 1980). There are now known to have been ten corrie glaciers in central Skye, in addition to two larger icefields (Walker et al. 1988; Ballantyne, 1989) whilst the Trotternish Peninsula in northern Skye supported two corrie glaciers (Ballantyne, 1990). The limits of the Loch Lomond Readvance in the Northern Highlands are rather less firmly established and this area is discussed in rather more detail below.

In the northernmost part of the Highlands, Sissons (1977a) identified evidence for sixty-seven former glaciers north of a line extending between Loch Torridon and the Cromarty Firth. Reconstructions of the extent of ice cover in this area indicate that glaciation during the Loch Lomond Stadial became increasingly restricted in extent towards the north (Robinson, 1977; Sissons, 1977a; Figure 2.7). In the extreme north few glaciers extend beyond their corries. Three main groups can be identified: those around Foinaven, Arkle and Meall Horn in the north; a second group in the vicinity of Ben More Assynt and Glen Oykel; and a third around Ben More Coigach (Sissons, 1977a; Lawson, 1986).

Farther south, in Wester Ross, Sissons' mapping of the Loch Lomond Readvance suggest that glaciers were largely restricted to mountainous terrain and were more widespread towards the west (Figure 2.8.). Ballantyne et al. (1987) identified seven centres of ice accumulation and dispersal in this area. The largest of these has been described as a small icefield or transection glacier complex, that developed between Lochs Maree and Torridon in the Torridon Hills (Sissons, 1977a; Ballantyne, 1986a) and an ice mass of similar dimensions is thought to have accumulated in the mountainous area between Glen Torridon and Strathcarron (Robinson, 1977). An extensive ice cap was also present on the Beinn Dearg massif (Reed, 1988). Further smaller readvance icefields have been identified in the Applecross Hills (Robinson, 1977) and the mountainous area between Loch Maree and Loch na Sealga (Sissons, 1977a), both of which fed small outlet glaciers. The An Teallach massif supported only small corrie glaciers (Sissons, 1977a) and in the Fannichs, several small corrie glaciers are believed to have existed, two of which coalesced
to form a large valley glacier (Sissons, 1977a; Reed, 1988). The limits of an icefield extending from the Torridon to the Rannoch area has recently been published (Bennett and Boulton, 1993a, 1993b), but is only approximate for much of its extent.

Figure 2.7 The Distribution of Loch Lomond Readvance glaciers in the Northern Scottish Highlands

Between Wester Ross and the Great Glen Fault the extent of Loch Lomond Readvance glaciation is not firmly established. Such limited evidence as has been published limits suggests that the western Grampians icefield extended north-west of the Great Glen but was diminishing in size (Thorp, 1986). Further limits to Loch Lomond Readvance glaciers have also been identified in areas to the north-west of the Great Glen (e.g. Sissons, 1967a; Peacock, 1970). The precise dimensions of this icefield are unknown, and there is almost no reliable information concerning the extent of Loch Lomond Readvance glaciation in the vast area of southern Ross-shire between Glen Moriston in the south-east (Sissons, 1979b) and Strath Carron in the north-west (Robinson, 1977). One of the principal aims of the research reported in this thesis is to establish the pattern and extent of Loch Lomond Readvance glacier ice in this hitherto-uninvestigated area (see section 3.3. below).
2.4. Other aspects of environmental change during the Late Devensian.

2.4.1. Introduction.

Late Devensian glaciation was accompanied by various other important environmental changes. Geomorphologically, perhaps the most important of these were periglaciation, sea-level change and slope development. The nature of these changes and their effects on Late Devensian (and Holocene) landscapes is discussed in turn below.

2.4.2. Periglaciation.

2.4.2.1.

The existence of periglacial landforms, deposits and structures in Scotland has long been recognised. Early publications written by the Officers of the Geological Survey included only brief references to such features (e.g. Harker, 1901; Crampton, 1911; Peach et al., 1913a, 1913b; Crampton and Carruthers, 1914) but research in the 1950s and 1960s (e.g. Galloway, 1958, 1961a, 1961b) provided more detailed (although still largely descriptive) accounts of periglacial phenomena throughout both upland and lowland Scotland. Research in the realm of periglacial studies has, however, accelerated rapidly in the last fifteen years. This recent work has highlighted the nature and range of periglacial landforms and deposits in Scotland in a much more systematic manner, by building on an increased understanding of the mechanisms of periglacial processes. This has also enabled assertions to be made regarding the age of periglacial features and controls on their distribution. The palaeoclimatic significance of such phenomena is also becoming more widely appreciated, and their implications for the reconstruction of glacial limits recognised.

It has long been established that relict periglacial structures (such as ice wedge casts and cryoturbation features) and periglacial deposits (particularly solifluction and head deposits) exist in lowland drifts. Recent work has also confirmed that a wide range of periglacial landforms, deposits and structures also exist on Scottish mountains. It has been suggested that almost all such upland features may be assigned to one of two mutually exclusive age categories: relict phenomena that were last active during the Late Devensian as a result of intensive periglacial activity, and features that developed during the Holocene and are active under the present-day 'maritime' periglacial environment of high mountain areas in Scotland (e.g. Sissons, 1979a; Ballantyne, 1984, 1987). Some Lastglacial features may have been reactivated during the Holocene. Although it has proved somewhat difficult to determine the age of periglacial phenomena on Scottish mountains, a number of approaches have been employed in an attempt to differentiate periglacial features of Late Devensian age from those that developed during the Holocene (Ballantyne, 1984). These approaches include pollen analysis and radiocarbon dating of
organic material buried under periglacial deposits, the measurement of present-day periglacial activity, and assessment of the present appearance of features as a guide as to whether they are active or relict. Perhaps the most powerful device for discriminating between Late Devensian and Holocene periglacial phenomena on Scottish mountains has been comparison of the distribution of such phenomena with that of glaciers that formed during the Loch Lomond Stadial, as periglacial features that occur within the limits of these glaciers presumably formed under the relatively mild conditions of the Holocene (Ballantyne, 1984). Table 2.1 summarises the range of Late Devensian and Holocene periglacial features on Scottish mountains.

### Table 2.1
The range of Late Devensian and Holocene periglacial features on Scottish mountains

<table>
<thead>
<tr>
<th>Category</th>
<th>Features of Late Devensian Age</th>
<th>Features of Holocene Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frost-weathered regolith</td>
<td>1. Mountain-top detritus (blockfields, stone pavements and debris surfaces)†</td>
<td>1. Some talus slopes (especially inside the limits of Loch Lomond Stadial glaciers)</td>
</tr>
<tr>
<td></td>
<td>2. Blockslopes and debris-mantled slopes†</td>
<td>2. Debris flows</td>
</tr>
<tr>
<td>Rapid mass-movement features</td>
<td>1. Most talus slopes (especially outside the limits of Loch Lomond Stadial glaciers)</td>
<td>3. Shallow translational slides</td>
</tr>
<tr>
<td></td>
<td>2. Protalus ramparts</td>
<td>4. Avalanche boulder tongues (?)</td>
</tr>
<tr>
<td></td>
<td>3. Avalanche boulder tongues (?)</td>
<td></td>
</tr>
<tr>
<td>Slow mass-movement features</td>
<td>1. Rock glaciers</td>
<td>1. Solifluction sheets and lobes</td>
</tr>
<tr>
<td></td>
<td>2. Boulder sheets and lobes</td>
<td>2. Ploughing boulders</td>
</tr>
<tr>
<td></td>
<td>3. Debris sheets and lobes</td>
<td>3. Turf-banked terraces</td>
</tr>
<tr>
<td></td>
<td>4. Solifluction sheets and lobes</td>
<td></td>
</tr>
<tr>
<td>Frost-sorting and mass-displacement</td>
<td>1. Large sorted circles</td>
<td>1. Small sorted circles</td>
</tr>
<tr>
<td></td>
<td>2. Large sorted stripes</td>
<td>2. Small sorted stripes</td>
</tr>
<tr>
<td></td>
<td>3. Earth hummocks</td>
<td>3. Earth hummocks (?)</td>
</tr>
<tr>
<td>Wind action features</td>
<td>4. Nonsorted stripes</td>
<td></td>
</tr>
<tr>
<td>Nival and fluvial landforms</td>
<td>1. Nivation hollows</td>
<td>1. Nivation hollows</td>
</tr>
<tr>
<td></td>
<td>2. Colluvial cones (?)</td>
<td>2. Colluvial cones</td>
</tr>
<tr>
<td></td>
<td>3. Alluvial fans</td>
<td>3. Alluvial fans</td>
</tr>
</tbody>
</table>

* Based on Ballantyne (1981) Table 14.2.
† Locally modified by Holocene frost sorting.
‡ Locally subject to intermittent movement under present-day conditions.
(?) Indicates uncertainty.

2.4.2.2. Late Devensian periglacidation.
A wide variety of Late Devensian periglacial landforms are known to have developed on Scottish mountains both before c. 13 ka BP and during the Loch Lomond Stadial of c. 11 - 10 ka BP. Almost all of these are now inactive. It is probable that the periglacial landscape of many mountain summits began to form as such summits emerged from beneath the last ice sheet as it decayed, but it seems likely also that a final period of intense periglacial activity during the Loch Lomond Stadial produced many of the large-
scale periglacial landforms now visible on high ground (Sissons, 1967a, 1983c; Price, 1983). Recent work by Ballantyne (1984, 1990) and Reed (1988) has also revived Godard's (1965) notion that certain summits in northern Scotland may have remained as nunataks above the level of the last ice sheet, and hence been subject to severe periglacial conditions over a much longer timescale.

Ballantyne (1984) has shown that the dominant characteristic of Late Devensian periglacial weathering on Scottish mountains was the production of three types of mountain-top detritus (openwork block deposits, sandy diamicts and silt-rich frost-susceptible diamicts (Figure 2.8), each of which supports a characteristic assemblage of relict landforms resulting from mass movement and frost-sorting. Large-scale sorted circles and stripes, earth hummocks and non-sorted relief stripes, sorted and non-sorted solifluction features, blockfields, massive boulder sheets and lobes and nivation benches are known to have developed on upper slopes before the end of the Loch Lomond Stadial (e.g. Kelletat, 1970; King, 1971; Sissons, 1976a; Shaw, 1977; Ballantyne, 1981, 1984). Following ice-sheet deglaciation, talus accumulated rapidly at the foot of rock slopes exceeding 40° (Ballantyne and Eckford, 1984). During the Loch Lomond Stadial protalus ramparts (Ballantyne and Kirkbride 1976) and protalus rock glaciers (Dawson, 1977; Sissons, 1979b) formed at the base of some talus slopes. Some avalanche boulder tongues and alluvial fans may also have formed under severe periglacial conditions at this time (Ballantyne, 1984).

Figure 2.8

Types of mountain top detritus on Scottish mountains
2.4.2.3. Holocene periglaciation.

Whereas the characteristic assemblage of Late Devensian periglacial landforms consists of features formed by intense cold and deep ground freezing, Holocene landforms are limited to those formed under the present 'maritime' periglacial climate, which is dominated merely by extreme wetness and exposure rather than severe cold or deep ground freezing (Ballantyne, 1987). Holocene frost action landforms are characteristically small and developed only to relatively shallow depths. Active frost action forms include small sorted circles and stripes and solifluction lobes active to depths of only 0.5 m or less (Ballantyne, 1987). Conversely, extreme wetness favours ploughing boulder movement and strong winds promote the formation of niveo-aeolian deposits and other aeolian features on certain lithologies, such as sandstone and granite. Recent years have witnessed a growing appreciation of the nature and frequency of snow avalanches in parts of the Scottish Highlands (Ward, 1984). Although most Scottish snow avalanches accomplish only limited geomorphic activity (Ward, 1985; Davison and Davison, 1987), examples of recent avalanche boulder tongues have been identified in the Cairngorms and eastern Grampians and avalanche impact landforms have been shown to occur on Ben Nevis (Ballantyne, 1989b).

2.4.2.5. Significance of periglaciation to environmental change.

The possibilities of distinguishing between landforms and deposits of different ages has implications for the identification of the limits of former glaciers, particularly those of the Loch Lomond Readvance. Certain relict (Late Devensian) periglacial features consistently extend up to limits of former glaciers but are absent within these limits, and hence are diagnostic of ground that escaped glaciation during the Loch Lomond Stadial. Periglacial deposits indicative of terrain that escaped glaciation during the Loch Lomond Stadial include intensely frost-shattered rock and mountain top detritus, together with gelifluction sheets and lobes of coarse bouldery detritus (Thorp, 1981; Ballantyne, 1984). Large scale patterned ground, protalus ramparts and rock glaciers also appear to be restricted in distribution to areas outside the limits of the Loch Lomond Readvance (Sissons, 1979a; Ballantyne, 1984; Ballantyne and Kirkbride, 1986). Similarly, a contrast in the degree of frost-shattering and frost-wedging of bedrock inside and outside the limits of the Loch Lomond Readvance has been noted, for example in Applecross (Robinson, 1977) and in the corries of An Teallach (Ballantyne, 1982).

Furthermore, recent research on the distribution of thick, in situ mountain-top detritus in the Northern Highlands strongly suggests that the upper limit of the Late Devensian ice sheet may be defined by the lower limit of such detritus (Ballantyne, 1984; Ballantyne et
In this area, the transition with increasing altitude from glacially-scoured bedrock to slopes mantled by a thick cover of frost-shattered detritus is not gradual (as would be expected if climate was the only control on the altitudinal limits of such detritus) but, on most lithologies, occurs over a distance of few tens of metres. This strongly suggests that the boundary represents the upper limit of a former ice surface, and Reed (1988) has argued that this former ice surface represents that of the Late Devensian ice sheet at its maximum extent. If this is the case, then many relict periglacial features may have developed throughout the Devensian on nunataks that protruded through the last ice sheet and possibly earlier ice sheets also. Much of the deep frost-shattered detritus on mountains in the Northern Highlands may have formed under severe periglacial conditions throughout much of the Late Devensian or longer. This is unlikely, however, to apply to all Scottish mountains. The last ice sheet was apparently more widespread and of greater altitude in the area of the Grampians and Southern Uplands (Sissons, 1981, Thorp, 1987) and the formation of in situ frost-shattered detritus in these areas may well have taken place only during the initial (cold) stages of ice-sheet downwastage and during the Loch Lomond Stadial. Moreover, although many of the periglacial landforms that developed on this detritus may have been initiated during ice sheet decay, the forms now visible are probably the products of the severe climate of the Loch Lomond Stadial (Sissons, 1983; Ballantyne, 1984).

On present evidence, however, it appears that it is impossible to attribute most Late Devensian periglacial phenomena to formation during the Loch Lomond Stadial alone. The occurrence of such phenomena directly outside the limits of the Loch Lomond Stadial glaciers has been interpreted by some authors (e.g. Gray and Lowe, 1977) as implying contemporaneity with Loch Lomond Stadial glaciation. This is inconclusive, as it is possible that stadial glaciers may have advanced into and removed previously-developed periglacial features (Ballantyne, 1984). Only certain valley floor forms such as rock glaciers, protalus ramparts and relict talus may be unequivocally assigned a Loch Lomond Stadial age as the mode of formation of such features required a renewal of stadial conditions following the decline of the last ice sheet (Ballantyne, 1984; Ballantyne and Kirkbride, 1986). Relict periglacial phenomena of Late Devensian age nevertheless have implications for the reconstruction of palaeoclimates. This topic is considered in section 2.5 below.

2.4.3. Late Devensian sea-level change.

Much research over the last twenty years has been devoted to establishing the pattern of Late Devensian sea level change in Scotland. The complex inter-relationship of sea level, responding eustatically to changes in global ice volumes together with a local Scottish ice mass advancing and retreating over coastal areas, and a land mass responding
differentially to isostatic recovery, has produced a complex series of displaced shorelines around the Scottish coast. There is evidence of both a morphological and stratigraphical nature for former sea-level changes in Scotland. A number of geomorphological methods, including large-scale geomorphological mapping of coastal areas, measurement of the altitude of coastal features and biostratigraphic analyses of buried peats and marine sediments have revealed evidence for an altitudinally distinct series of shorelines.

Evidence for former sea levels may be studied with a view to establishing a chronology of glaciation of an area. Relative movements of land and sea have in effect been determined by the advance and retreat of glacier ice, and the investigation of raised shorelines may substantiate the age and chronology of glacial features. The absence of a particular shoreline may be the result of ice occupying the area at the time the shoreline was formed (Gray, 1985). Alternatively its absence may be due to destruction by an ice advance subsequent to its formation.

2.4.4. Late Devensian palaeoclimate

2.4.4.1. Introduction.
Quantitative estimates of climatic parameters for the Late Devensian period have been provided by both geomorphological and biological evidence. Few palaeoclimatic inferences have been made with regard to the Dimlington Stadial but there is rather more information concerning the Lateglacial Interstadial and the Loch Lomond Stadial. Palaeoclimatic information for these three chronozones is summarised below.

2.4.4.2 The climate of the Dimlington Stadial.
There is only a very limited amount of data concerning the climate of the Dimlington Stadial in Scotland. Sutherland et al., (1984) estimated a mean July temperature of 4°C for the Late Devensian glacial maximum on St. Kilda on the basis of the equilibrium line altitudes of contemporaneous glaciers. Geomorphological evidence confirms that temperatures were low, since permafrost is known to have existed on low ground during the initial stages of ice sheet retreat, as demonstrated by the presence of certain fossil ice wedges (Sissons, 1980b). Precipitation estimates for Britain during ice-sheet build-up vary but values approaching 150% of that of the present day have been suggested for major accumulation and dispersal areas (Manley, 1975). A value of 700 mm has been proposed for North East Scotland from modelling the general circulation (Williams and Barry, 1974).

It has been suggested that the initial downwastage of the last Scottish ice sheet occurred not as a result of warming, but because of a decrease in snowfall. This seems reasonable, as the southerly migration of the North Atlantic oceanic polar front during the Dimlington
Stadial left the British Isles surrounded by cold polar water (Ruddiman and McIntyre, 1975, 1981; Lamb, 1977; Ruddiman et al., 1977; Sissons, 1981, 1983).

2.4.4.3 Climate during the Lateglacial Interstadial.
Studies of both marine and terrestrial faunal assemblages have proved instrumental in obtaining an insight into the climate of the Lateglacial Interstadial. Peacock (1981b; 1983) proposed a complex history of Lateglacial Interstadial climate from his analysis of mollusc and foraminiferal assemblages in the Clyde Beds. His suggestion that there was an initial period of particularly mild conditions is supported by research on coleoptera assemblages on land (Bishop and Coope, 1977; Coope, 1977). Between 12.5 and 12 ka BP there appears to have been a cooling with summer surface temperatures on land probably 1-2° less than those of the present. Firmly dated evidence from boreholes off the west coast indicates a marine climate some 2-3° lower than the present for much of the interstadial (Peacock, 1983). The surface water marine record suggests that a short period of milder conditions occurred at c. 11 ka BP (Peacock, 1981b, 1983). This contrasts with the terrestrial record, which indicates declining temperatures at this time, a situation that would have been particularly conducive to glacier growth in the early part of the Loch Lomond Stadial (Sissons, 1980b). Such conflicts are, however, dependent on radiocarbon dating. A marked contrast in oceanic temperatures during the Lateglacial Interstadial between the Scottish west coast and southern Norwegian sea has also been noted (Sutherland, 1984a). This implies a marked climatic gradient across northern Scotland and it appears likely that the west - east climatic contrasts were also considerably greater than those of the present.

2.4.4.4 The climate of the Loch Lomond Stadial.
Climatic data for the Loch Lomond Stadial have been derived from four sources of evidence: the dimensions of Loch Lomond Readvance glaciers at their maximum extent; climatic implications of relict periglacial phenomena; evidence afforded by coleoptera; and evidence provided by contemporaneous pollen assemblages. The information derived from each of these sources is considered in turn below.

The distribution of Loch Lomond Readvance glaciers and variations in their extent and altitude have been widely used as a basis for palaeoclimatic inference and calculation. Data relating to glacier size, potential snow blowing areas, potential avalanche areas, insolation receipt and effect of shadow (c.f. Sissons and Sutherland, 1976; Sutherland, 1984b) have all been employed in determining palaeoclimatic conditions, but the most important palaeoclimatic parameter is the reconstructed equilibrium line altitude or ELA of former glaciers (Sissons, 1974b). Research on the mass balance of modern glaciers has revealed that the ELA provides a critical link between glaciers and climate (Sutherland,
(1984b), as this parameter represents the altitude on the surface of a glacier where accumulation and ablation are exactly balanced when the glacier is in equilibrium. For reconstructed glaciers, this implies that if glacier A had a higher ELA than glacier B, then A must have experienced either less snow accumulation or a greater rate of ablation or both of these. It follows that patterns of former ELAs across an area may be interpreted in terms of palaeoclimatic controls on glacier development. The altitude of equilibrium lines therefore enables inferences to be made with regard to several climatic variables including the former direction of snow-blowing winds and controls on regional snowfall. In addition, it has been demonstrated using modern glacier analogues in Norway that mean summer temperatures can be related to glacier variables such as the ELA (Liestøl, 1967) and that there is a close relationship between mean summer temperature and accumulation at the equilibrium line (Sutherland, 1984b). It is therefore possible to make inferences concerning summer temperature for given or assumed values of accumulation.

A number of researchers have provided information about the climate during the stadial from reconstructions of the dimensions of Loch Lomond Readvance glaciers mapped throughout Scotland. (e.g. Sissons, 1974b; Sissons and Sutherland, 1976; Sissons, 1980b; Ballantyne, 1989). The local distribution of Loch Lomond Readvance glaciers across particular mountain groups and the associated pattern of ELAs has been employed to derive information concerning the impact of climatic variables on the formation of this pattern. In an analysis of the Loch Lomond Readvance in the northern mainland of Scotland, for example, Sissons (1977a) noted a general inland rise in ELAs. He suggested that although snowfall diminished inland from the west coast there were local deviations from this trend apparently due to variations in the amount of snow blown on to individual glaciers. In Wester Ross, a general west to east rise in ELAs and an accompanying decrease in the intensity of glacierisation has been noted (Ballantyne et al., 1987). Ballantyne (1989) also described a local decline in ELAs towards the north east across the former Cuillin icefield on Skye, and interpreted this in terms of an easterly transfer of snow across ice sheds by westerly winds, although he also suggested that the ELA of former corrie glacier indicate that the dominant snow-bearing winds were southerlies.

The SE Grampians have been analysed in relation to the areas from which snow could have blown on to the glaciers from the four major quadrants. This comparison has shown that southerly winds were the most important in the transfer of snow on to glacier surfaces (Sissons and Sutherland, 1976; Sissons, 1980b). In synoptic terms this situation has been explained by the dominance of precipitation associated with southerly or easterly winds with accompanied snow blowing as fronts crossed the mountains and winds veered towards the SW (Sutherland, 1984b).
With regard to insolation and the effect of shadow it has been noted that in some areas the largest valley glaciers had a southerly rather than a northerly aspect (Sissons, 1977a, 1977b, 1979a; Sissons and Sutherland, 1976; Sutherland, 1984b). It has been inferred from this that there was typically considerable summer cloudiness during the stadial so that north-south contrasts in direct insolation during the ablation season were minimised.

The ELAs of Loch Lomond Readvance glaciers have also been employed to reconstruct regional palaeoclimatic patterns across the entire Scottish Highlands (Sissons, 1980b; 1983c; Figure, 2.9). From these maps it is apparent that ELAs rose significantly both eastwards, away from the west coast and northwards from the Highland Boundary into the area of generally higher ground. From this generalised distribution of ELAs and the contrast in the extent and volume of glaciers in the South West Highlands and Cairngorms, Sissons (1980b) inferred that the principal control on differential glacierisation was the pattern of precipitation. He emphasised the fact that there was probably very low precipitation in the Cairngorms and Monadhliath mountains and estimated that the Cairngorm summits experienced precipitation of perhaps as little as 500 - 600 mm yr\(^{-1}\). He also suggested that contemporaneous annual precipitation in the adjacent Spey valley may have been as low as 200 - 300 mm yr\(^{-1}\). In contrast with the figure for the Cairngorms, it was calculated that precipitation during the Loch Lomond Stadial in the South East Grampians was 1000 - 1500 mm (Sissons and Sutherland, 1976).

**Figure 2.9**

*The pattern of equilibrium line altitudes across Scotland*

*(Sissons, 1977)*
In the Western Grampians, ELAs displayed an overall rise eastwards from below 400 m to above 600 m, indicating a decrease in snowfall with distance from the west coast. Superimposed on this trend, however, was a northwards rise in ELAs which Sissons (1980b) attributed to heavier snowfall in the South West Grampians than the North-West Highlands. He inferred that this trend reflected that of winter depressions following a more southerly track than at present. As a result, glaciers extended to low altitudes (sometimes to sea level) in the western Grampians and the islands of the Inner Hebrides such as Mull and Skye (Gray and Brooks, 1972; Ballantyne, 1989). Glaciers further north, in Assynt, however, tended to be small and confined to mountain areas (Sissons, 1977a; Lawson, 1987). It should be noted, however, that although the general pattern of ELAs reconstructed by Sissons (Figure 2.9) is useful in indicating the pattern of snowfall distribution during the stadial, derivation of absolute values for stadial precipitation has so far proved impossible.

Calculation of temperatures during the Loch Lomond Stadial has also been attempted. Mean summer temperatures have been inferred using relationships established between average accumulation and mean ablation season temperature for modern Norwegian glaciers (Sutherland, 1984b). Sissons and Sutherland (1976), for example, calculated a mean summer temperature of c. 4.5°C for the South-East Grampians, which implies a stadial July sea-level temperature of c. 6°C. Similarly a mean July sea-level temperature of c. 7°C has been estimated for the South-West Grampians on the basis of the dimensions of stadial glaciers (Sissons, 1976b) and Ballantyne (1989b) employed the same method to calculate a mean July temperature c. 6°C for the Isle of Skye.

The available palaeoclimatic data imply a stormy climate with frequent depressions tracking across Scotland (Sutherland, 1984a). Sissons and Sutherland (1976) suggested that this was related to the position of the oceanic polar front, which, being located off the coast of the British Isles for much of the Loch Lomond Stadial, is believed to have both generated and directed depressions. This suggestion has received support from the deep sea core studies of Ruddiman et al. (1980) and Duplessey et al. (1981).

Relict periglacial features formed under the cold conditions of the Late Devensian also provide information regarding the nature of the climate at the time of their development (Ballantyne, 1984). Most of the large-scale periglacial landforms that exist in Scotland are relict forms and can be distinguished from Holocene forms using the criteria described in section 2.4. (see above), although there have been some difficulties in distinguishing features formed during the downwasting of the last Scottish ice sheet (or earlier) from those that developed during the Loch Lomond Stadial.
Studies in present-day periglacial environments have shown that certain periglacial features form only under permafrost conditions and within a certain temperature range. This means that the occurrence of similar features in relict form in Scotland may be employed as evidence of former permafrost and, more contentiously, palaeotemperature. The widespread distribution of former permafrost down to sea level in Scotland is indicated by features such as ice wedge casts (Galloway, 1961; Sissons, 1974a; Watson, 1977). The date of such ice wedge casts, however, is not often firmly established. Although most casts occur outside the limits of the Loch Lomond Readvance glaciers and may have formed during the stadial, it is also possible that they may have developed during the earlier stages of ice sheet retreat or in some cases even earlier, outside the maximal limits of the Late Devensian ice sheet. The palaeoclimatic implications of their formation are more certain. Research on modern ice wedges (in Alaska) has shown that such features are restricted to areas of continuous permafrost, which implies that mean annual temperatures, not exceeding -5°C to -8°C are required for wedge formation (Péwé, 1966; Brown and Péwé, 1973; Washburn, 1979) though exceptionally, wedges may develop in silts where mean annual temperatures are as high as -3.5°C (Hamilton et al., 1983). Thus the presence of the Scottish ice wedge casts may indicate that mean annual temperatures were no higher than -5°C to -8°C at the time of their formation, which may or may not have coincided with the Loch Lomond Stadial. The reported occurrence of three ice wedge casts in Loch Lomond Stadial sediments would seem to imply continuous permafrost conditions during the Loch Lomond Stadial (Sissons, 1974a, 1976a, 1976b, 1980b), but as this evidence is sketchily documented, some doubt remains as to its significance (Ballantyne, 1984).

Relict rock glaciers provide further information regarding the climate of the Loch Lomond Stadial in Scotland. These rock glaciers are considered to be indisputably of Loch Lomond Stadial age (Ballantyne, 1984), although their thermal implications are not so clear. It is generally accepted that rock glaciers may form under conditions marginal for permafrost (or where there is only localised permafrost) under mean annual air temperatures of up to 0°C (Corte, 1976; Washburn, 1980). The distribution of rock glaciers in Scotland implies that mean annual air temperatures during the stadial were no higher than 0°C at 350 - 400 m and therefore did not exceed c. 2°C at sea level, figures that must, however, be considered to be conservative estimates (Ballantyne, 1984).

Similarly, large-scale patterned ground is widely held to be strongly indicative of former ice-rich permafrost. Although climatic controls on such features are not yet fully understood, Goldthwait's (1976) suggestion that forms > 2 m in diameter develop in continuous or slightly discontinuous permafrost where mean annual temperatures have
been -4°C to -6°C or lower 'for decades or centuries' provides some indication of former temperature regimes. These features cannot be readily dated however, and are not necessarily of Loch Lomond Stadial age.

In sum, the above evidence indicates that much of Scotland experienced continuous permafrost down to sea level, at some time subsequent to the retreat of the Late Devensian ice sheet, with associated mean annual air temperatures no higher than -5°C and possibly -8°C to -10°C or less (Ballantyne, 1984). Assuming that some ice wedge casts are indeed of Loch Lomond Stadial age and also assuming that such casts imply a mean annual air temperature of -5°C or less and that the estimated ELAs of contemporaneous glaciers are correct, Ballantyne (1984) suggested that mean January sea level temperatures must have been no higher than c. -17°C in the Western Grampians under full stadial conditions and at 600 m and 1000 m would have been no higher than -20°C and -23°C respectively. The validity of these estimates, however, rests heavily on the rather equivocal evidence for ice wedge cast formation during the Loch Lomond Stadial.

Only broad generalisations can be made from periglacial evidence about former precipitation patterns, as the controls of precipitation on most periglacial landforms are largely unknown. However, an eastwards rise in the lower limit attained by relict boulder sheets and lobes and large scale sorted patterned ground may reflect an eastwards decrease in effective precipitation at the time of formation (Ballantyne, 1984), a trend consistent with the eastwards rise in altitude of Loch Lomond Stadial protalus ramparts (Ballantyne and Kirkbride, 1986) and the reconstructed ELAs of contemporaneous glaciers. In comparison with the western Highlands, precipitation in the Cairngorms during the Loch Lomond Stadial appears to have been rather low. This evidence of former aridity is confirmed by the concentration of relict rock glaciers here, as such features are indicative of moderate or light snowfall (Thompson, 1962; Blagborough and Farkas, 1968).

The glacial and periglacial evidence discussed above therefore indicates marked precipitation contrasts across Scotland during the Loch Lomond Stadial, although estimates of the absolute values involved are highly generalised. This trend is apparently supported by evidence from pollen assemblages. There is a clear inverse relationship between areas of high precipitation indicated by the above mentioned evidence and the distribution of pollen sites in which the stadial spectra contain high frequencies of *Artemisia* (Pennington et al., 1972; Walker, 1975a, 1975b, 1977; Lowe, 1978; Lowe and Walker, 1977; Birks and Mathewes, 1978). Initial suggestions of spatial climatic variability in the Loch Lomond Stadial came from Birks and Matthews (1978), who examined the percentage representation of *Artemisia* at Scottish Late Devensian pollen
sites and argued that aridity was induced by a rainshadow effect from upland areas. Sissons (1980) likewise considered that high *Artemisia* percentages reflected relatively arid conditions and highlighted the fact that for sites in the Scottish Highlands, relatively high percentages of *Artemisia* pollen were associated with higher regional ELAs which indicates that both reflected response to precipitation totals across Scotland. Several researchers have also suggested that there was a mid-Loch Lomond Stadial increase in aridity (Caseldine, 1980; Macpherson, 1980; Tipping, 1984). The evidence for this is based largely on the temporal distribution of high *Artemisia* percentages during the stadial. A study of the distribution of *Empetrum* and *Salix* at the same horizon as the *Artemisia* maximum also supports the conclusion that the northern and eastern Highlands were much drier than the western and southern Highlands (Macpherson, 1980).

The lack of modern analogues for the vegetation communities that existed during the Lateglacial, together with the relative slowness of response of vegetational changes to improving climatic conditions make direct climatic inferences from pollen data very difficult (Pennington, 1977b). Temperature changes can be much more readily inferred from coleoptera (Atkinson *et al.*, 1987) although terrestrial faunal evidence related to the Loch Lomond Stadial in Scotland is rather limited.

Coleopteran assemblages from sites in South-West Scotland indicate a climatic deterioration at the beginning of the stadial. On the basis of coleopteran evidence Bishop and Coope (1977) constructed a temperature curve for the Loch Lomond Stadial in this area with average July temperatures falling from 12°C at c. 11 ka BP and continuing to decline to below 9°C at c. 10.2 ka BP, before rising steeply to reach 15°C by 9.7 ka BP. These figures are in reasonable agreement with those calculated on the basis of inferred ELAs for Loch Lomond Readvance glaciers.

### 2.5. Late Devensian vegetation development

#### 2.5.1.

The development of vegetation during the Lateglacial and Postglacial period has been reconstructed partly from macrofossil evidence, but mostly from evidence from pollen stratigraphic studies. Attempts have been made to correlate Late Devensian biostratigraphic units and boundaries across Scotland (e.g. Pennington, 1977) but regional differences in Late Devensian vegetation development have frustrated this. Some regional pollen chronostratigraphies, however, have been established (e.g. Pennington, 1975a). The general pattern of vegetation change throughout Scotland for the period following the withdrawal of the Late Devensian ice sheet is outlined briefly below and a
final section describes in somewhat greater detail, the nature of vegetation development in the North West Highlands.

2.5.2. Late Devensian Lateglacial Interstadial

Over seventy Lateglacial pollen sites have been published to date in Scotland. Their distribution is shown in Figure 2.10. The principal features of Lateglacial vegetation development and regional variations, as revealed by these sites and reconstructed from pollen analysis, are highlighted below.

**Figure 2.10**
The distribution of lateglacial pollen sites in Scotland
Evidence from pollen diagrams has indicated that distinct pioneer vegetation communities colonised newly deglaciated mineral substrates. This initial phase of colonisation is thought to have occurred soon after the recession of the last ice sheet, as a number of basal sediments containing pollen evidence of such pioneer species have yielded radiocarbon dates of c. 13 ka BP (Kirk and Godwin, 1963; Sissons and Walker, 1974; Vasari, 1977; Walker and Lowe, 1982).

At the base of many Lateglacial pollen profiles in Scotland a period of low pollen production is apparent (Gray and Lowe, 1977). The characteristic pollen taxa at these levels include Gramineae, Cyperaceae, Caryophyllaceae, Chenopodiaceae, Compositae, Artemisia, Rumex, Salix and Lycopodium, species indicative of an open habitat. A species rich grassland is known to have developed throughout Scotland. Low frequencies of woody plant pollen at the base of the pollen profiles have been noted, suggesting only a sparse presence of shrubs and trees.

Such arboreal grains as have been recorded, consisting principally of Betula and Pinus, are widely held to be of long distance provenance (Lowe and Walker, 1977) (although Betula nana may have been local in origin; c.f. Andersen, 1981; Kolstrup, 1986). It has been suggested, however, that their occurrence may be attributed, to some degree, to the reworking of older sediments around basins (Cundill and Whittington, 1983).

Following the pioneer stage, the further development of vegetation became more complex. Various environmental factors such as altitude, aspect, exposure, soil conditions and edaphic variations are considered to have caused differences in vegetation development between sites (Walker, 1984). In general terms, this notably treeless, species-rich grassland was succeeded by a widespread dwarf-shrub heath with variable amounts of Empetrum and Salix and localised copses of tree birch.

The differences in vegetation development due to environmental factors has been demonstrated in the Grampians. In lower and more sheltered areas, as recorded in the south east Grampians, for instance, a closed vegetation cover developed (Lowe and Walker, 1977) consisting mainly of grassland, with juniper, dwarf birch, willow and some tree birch. A somewhat different pattern emerges from the evidence from sites at greater altitudes and in more exposed localities in this region, where moss heaths and poor grassland communities are believed to have developed (Walker, 1975b; Lowe and Walker, 1977; Lowe, 1978). A similar vegetation pattern is believed to have evolved in Fife (Whittington et al., 1990), the central lowlands (Donner, 1957; Newey, 1970) and in the Scottish borders (Webb and Moore, 1982). The characteristic Lateglacial interstadial vegetation cover in the south west (Moar, 1969) and western seaboard (Rymer, 1977) was of local stands of juniper, Empetrum and Salix, and tree birch in places.
Further north in the valleys of the Central Grampians a succession from the pioneer grass- and sedge- dominated communities to one of shrub tundra' (Birks and Mathewes, 1978) consisting primarily of *Empetrum*, with some *Betula* and *Salix* was described (Walker, 1975a). In Aberdeenshire the trend for a greater expansion of *Empetrum* towards the more northerly areas has been demonstrated. Here, limited stands of both pine and tree birch are also recorded (Vasari and Vasari, 1968; Vasari, 1977). In north eastern Scotland and on Orkney and Shetland, dwarf shrub heath was widespread but here *Empetrum* appears to have been particularly important, although it was similarly interspersed with open grassland and a variable representation of juniper (Birnie, 1981; Moar, 1969b; Pennington *et al.*, 1972; Pennington, 1977b). This pattern also occurred in the North West Highlands although a virtual absence of tree birch during the Lateglacial Interstadial has been noted (Pennington, 1975). The vegetation succession of this area is considered in greater detail in section 2.5.4.

In most Scottish Lateglacial profiles there is evidence for a 'progressive unidirectional vegetation succession' (Walker, 1984; p. 385) from pioneer communities to grass, heath and shrub development in response to the initial rise in temperatures. From some sites, however, there are indications of a short-lived vegetational revertance phase during the Lateglacial Interstadial. This is believed to correspond to the Bölling - Older Dryas - Alleröd sequence identified in North West Europe. This sequence is represented typically by a decline in frequencies of woody plant pollen, principally *Empetrum*, *Juniperus* and *Betula*, and an associated increase in percentages of such open habitat taxa as *Rumex*, *Artemisia*, Caryophyllaceae and Gramineae. The observed lithostratigraphic change from organic to a more mineral sediment which accompanies this pollen taxa change appears to confirm this trend in the biostratigraphic sequence.

This change in pollen spectra reflecting a minor climatic recession has been identified in the south-east Grampians (Walker, 1977; Caseldine, 1980), from north-east Scotland (Vasari and Vasari, 1968; Vasari, 1977; Clapperton *et al.*, 1975) and Keith (Donner, 1957) and further south at Stormont Loch, Blairgowrie (Caseldine, 1980) and Fife (Whittington *et al.*, 1990). It is also well represented in north-west Scotland (e.g. Pennington, 1977; see below) where evidence for this period of increased soil erosion and an inferred short episode of more severe climatic conditions has been recognised in five profiles from lakes in this area. An age of 12 - 11.8 ka BP has been proposed for the revertance phase on the basis of radiocarbon dating of sediments from profiles displaying pollen evidence for this phase (Pennington, 1975, 1977a, b).
2.5.3. Loch Lomond Stadial.

The effect of the climate of the period of the Loch Lomond Stadial is well represented in pollen profiles throughout Scotland (Vasari and Vasari, 1968; Vasari, 1977; Pennington et al., 1972; Pennington, 1977, a, b; Walker and Lowe, 1977; Birks and Matthewes, 1978; Walker and Lowe, 1979; Caseldine, 1980; Macpherson, 1980). Lower pollen concentrations, poorer pollen and spore preservation and the reappearance of predominantly minerogenic sediments and decline of shrubs and trees imply a marked climatic deterioration (Gray and Lowe, 1977). Pollen stratigraphic evidence indicates that the plant communities present during the Loch Lomond Stadial are analogous to the present day tundra regions and shrub-dominated sub-alpine and low alpine zones of Scandinavia (Birks, 1973a). The shrub-dominated plant communities of the Loch Lomond Stadial, characteristic of bare or broken soils, included various herbaceous species such as Compositae, Caryophyllaceae, Cruciferae, Chenopodiaceae, Artemisia, Rumex/Oxyria and the clubmosses, Lycopodium and Selaginella. This is believed to reflect the dramatic break-up of the vegetation cover and destruction of the soils that had gradually formed during the Lateglacial Interstadial.

Some uniformity in Loch Lomond Stadial pollen assemblages has been noted although there are both regional and local variations, some of which, it is suggested, may be related to thickness and duration of snow cover (e.g. Pennington, 1977). In the Grampians, low Artemisia values in Loch Lomond Stadial sediments from the east have been attributed to extensive snow cover (Walker, 1975b) whilst high Artemisia counts for the central Grampians have been interpreted as reflecting a more arid climatic regime (Birks and Matthewes, 1978). Further, it has been proposed that increases in Artemisia percentages in Loch Lomond Stadial pollen assemblages in other profiles may be indicative of changes in both snow cover and precipitation levels during this period (Caseldine, 1980; Macpherson, 1980; Pennington, 1980).

2.5.4. Early Postglacial.

Rates of colonisation and vegetational change during the early Postglacial have been established with the radiocarbon dating of Flandrian sediments (Moar, 1969b; Nichols, 1967; Birks, H.H., 1970, 1973; Birks, H.J.B., 1973b; Pennington et al., 1972; Hibbert and Switsur, 1976; Birks and Matthewes, 1978; O'Sullivan, 1976; Peglar, 1979; Dickson, et al., 1978); Whittington et al, 1990). Sediments of early Flandrian age in most pollen stratigraphies from sites in Scotland are characterised by pollen spectra dominated by open habitat taxa, notably Gramineae and Rumex. These were succeeded by taxa
indicative of dwarf shrub and scrub vegetation in which, *Juniperus, Empetrum* and *Salix* were locally dominant. Subsequent horizons contain high frequencies of *Betula*, and *Corylus* pollen grains. This initial herbaceous phase is not recorded, however, in pollen stratigraphies from sites within the areas covered by Loch Lomond glaciers. Here, the earliest pollen spectra are dominated by dwarf-shrub taxa and especially by *Empetrum* (e.g. Walker and Lowe, 1977).

In many early postglacial pollen diagrams there is an early peak in the curve for *Juniperus communis*, which represents the replacement of herbaceous park tundra vegetation by juniper scrub. It is thought that it may also reflect the increased flowering, under ameliorating climate, of those junipers already present in the vegetation cover. A number of singular radiocarbon dates provides some indication of the development of juniper throughout Scotland. An extensive *Juniperus* cover was present by c. 10.3 ka BP in the Scottish border area and around the southern margins of the Grampian Highlands, (Hibbert and Switsur, 1976; Lowe, 1978), although it was not until 10.0 ka BP that it was widely established throughout the Grampian Highlands (Vasari, 1977; Walker and Lowe, 1979) and in Skye (Williams, 1977). Even later dates ranging from 9.7 - 9.5 ka BP have been recorded at sites in the Spey valley (Birks and Mathewes, 1978) and on Mull (Walker and Lowe, 1982). A recent study of the synchronicity of the *Juniperus* peak however, in which an statistical analysis of the distribution of such radiocarbon dates throughout Scotland was undertaken, has indicated that there may have been local stands of juniper independent of a regional trend (Tipping, 1986).

Following the spread of juniper, a deciduous woodland gradually progressed further northwards. An initial northwards expansion of *Betula* was subsequently succeeded by one of *Corylus avellana*. In southern Scotland and around the southern and western margins of the Grampians both *Betula verrucosa* and *Betula pubescens* were present. Further north, *B. verrucosa* appears to have been more widespread (Birks, 1977). *Sorbus aucuparia, Salix* spp., and *Populus tremula* were also present in the open birch woodland and local stands of *Juniperus communis* occurred in places, although juniper did not survive after the arrival of *Corylus avellana*. Radiocarbon dates suggest that tree birch had become established throughout central and southern Scotland shortly after 10 ka BP (Vasari, 1977; Lowe, 1977) and had become widespread in the extreme north of Scotland well before 9 ka BP (Peglar, 1979). *Corylus* is believed to have spread into lowland Scotland by 9.3 ka BP (Hibbert and Switsur 1976; Lowe, 1978) along the western seaboard and into Wester Ross by 8.8 ka BP (Birks 1972; Williams 1977; Walker and Lowe, 1982) and into the central Grampians by 8.7 ka BP. (Birks and Mathewes, 1970).
2.5.5. The pattern of Late Devensian vegetation development in North West Scotland.

The general vegetational development of north-west Scotland has been elucidated through a number of pollen-analytical studies (Kirk and Godwin, 1963; Vasari and Vasari, 1968; Moar, 1969; H.H. Birks, 1972; Pennington *et al.*, 1972; Birks, 1973; Pennington, 1975a, 1977a; Pennington *et al.*, 1977; Robinson, 1977; Lowe and Walker, 1986; Walker *et al.*, 1988). This work has been complemented by chemical and diatom studies of Late Devensian lacustrine sediments (Pennington, 1973, 1977a). A regional pollen zonation scheme has been proposed and on the basis of radiocarbon dating of sediments at Loch Droma, a Lateglacial and early Flandrian chronostratigraphy has been suggested (Pennington, 1975) for the North West Highlands (Figure 2.12).

A number of Lateglacial pollen sites have been investigated north west of the Great Glen Fault. Seven occur in Sutherland (Figure 2.12), the most important of which is Cam Loch, upon which a regional chronostratigraphy is based. On the Scottish mainland, only four other Lateglacial profiles have been identified from sites in Wester Ross in addition to those in Sutherland and no others are recorded between Applecross and the Great Glen fault. Although it was formerly believed that there were also a number of established Lateglacial sites on Skye (Birks, 1973) and that vegetation development on Skye was significantly different from that on the mainland, it is now considered that many of the sites contain only Flandrian pollen (Walker *et al.*, 1988). Additional sites which have recently been recognised on Skye (cf. Walker *et al.*, 1988) do have, however, Lateglacial sequences and therefore a comparison with the mainland regional pollen diagrams can be made more realistically.

**Figure 2.11**

A Lateglacial and early Flandrian regional pollen zonation scheme and chronostratigraphy (Pennington, 1977)
Pennington et al., (1972) analysed cores from a number of lochs in Sutherland and Wester Ross and were able to identify six Postglacial Regional Pollen assemblage zones (NSI - NSVI). On the basis of radiocarbon dates from Loch Clair and Loch Sionascaig, these pollen assemblage zones were designated as chronozones (NWI - VI) for northwest Scotland. The Lateglacial stratigraphy from these sites was analysed and initially divided into three zones: A (Pre-Interstadial), B (Interstadial) and C (Post-Interstadial). These zones were not, however, radiocarbon-dated. The three Lateglacial zones present at Cam Loch were sub-divided into seven pollen assemblage zones (Pennington, 1975). A further comparison of four percentage Lateglacial pollen diagrams (Pennington, 1977) resulted in the definition of three regional pollen assemblage zones for the northern mainland of Scotland A, B and C. These zones were based on those described for Cam Loch (Pennington, 1975) but whereas A was originally termed 'pre-Interstadial', the Lateglacial interstadial was now correlated with zones A2, A3, and B (Pennington, 1975) on the basis of radiocarbon-dating. It is, therefore, possible to compare the differences in
pollen profiles from Sutherland, Wester Ross and Skye with the regional pollen assemblage zonation scheme (Figure 2.13-14).

According to Pennington's (1975; 1977) regional pollen zonation scheme pre-Interstadial zone A is a *Rumex* zone subdivided into three periods dominated first by *Huperzia selago*, then *Empetrum* and finally *Artemisia*. This is interpreted as being indicative of pioneer open communities on skeletal soils. Zone B is dominated by *Empetrum* with some juniper. (At some of the sites on which the scheme is based, this is subdivided into (B1), an *Empetrum* maximum, a subsequent recession of *Empetrum* (B2) and a subsequent *Juniperus* maximum in B3). The base of zone B is equated with a zone at Loch Droma dated to 12870± 155 yr BP and this zone is believed to have witnessed a more closed vegetation community. Zone C is defined as the Post-Interstadial *Artemisia* zone. It consists of assemblages which include taxa indicative of open, disturbed soil conditions.

A *Rumex - Lycopodium (Huperzia)*selago zone forms the transition between the Lateglacial and Postglacial pollen zones. It precedes the rise of *Betula* at c. 8 - 9 ka BP that forms the lower boundary of NWIII, a birch-hazel zone. NWIV is associated with the rise of *Pinus* c. 7.9 ka BP. A comparison of the regional pollen stratigraphy and the vegetational successions from other sites in north west Scotland demonstrates the similarities and differences between sites.

The patterns of vegetation development as recorded in pollen profiles at the various sites in north west Scotland are broadly similar to the regional pollen assemblage zones proposed by Pennington (1977). Diagrams from Sutherland, Wester Ross and Skye are consistent with this but aberrations from the regional pollen profile do occur. The major differences are highlighted below.

At the main site in Sutherland, Cam Loch, a basal radiocarbon date indicates that the expansion of woody plants occurred c. 13 ka BP. Here there is also evidence in local pollen assemblage zone Cb of an oscillation, where *Juniperus* and *Empetrum* rise then decline as *Artemisia* increases. This is interpreted by Pennington (1972) as being climatically significant in relation to the Bölling-Older Dryas fluctuation. This is not widely reflected in other sites in Sutherland. Similarly, an additional pollen zone was recognised below A1 at this site, a zone that is not generally present elsewhere. Three other pollen profiles from Loch Sionascaig, Lochan an Smuraich and Loch Tarff, all in Sutherland, have been shown to contain zones A, B and C, which are subdivided in the same way. The only notable major difference is recorded at Loch Sionascaig where in A2, where there is an increase of woody plants in the *Rumex-Empetrum* zone, *Juniperus*
pollen is also a significant component. Moar (1969) analysed a pollen profile in the north of Sutherland and discerned a basal zone of Lateglacial character with very high *Empetrum*, Gramineae, Cyperaceae, Rumex and Pteridophytes.

Further differences are apparent in terms of the transition zone between the Lateglacial and Postglacial that occurs at only some sites. Such a zone is present between Lateglacial period C and Postglacial NWSI at both Loch Sionascaig and Cam Loch, where high values of Cyperaceae, Rumex, *Huperzia selago* and *Salix* have been recorded.

In Wester Ross, Loch Droma has long been recognised as an important Lateglacial site (Kirk and Godwin, 1963). The pollen diagram from this site also appears to fit the pattern described for northern Scotland, although it was not zoned. However, with more sites available for comparison, it is now clear that almost the entire Lateglacial period is represented in this pollen profile. The lowermost part of the sequence correlates with the *Rumex* regional pollen zone A: here the basal section is dominated by *Huperzia selago* and Filicales as well as possessing high values for *Rumex* and Artemisia. The middle section correlates with zone B and the top part of the section of pollen analysed, with the *Artemisia* zone C (Pennington, 1972). Within zone A, the fluctuations recognised at Cam Loch and correlated with the Bölling-Older Dryas oscillation of North-West Europe chronostratigraphy, are also present at Loch Droma. The radiocarbon date of 12810 ± 155 yr BP at this site is from the A/B boundary, which has been dated elsewhere (Pennington, 1975) as c. 11.8 ka BP which is inconsistent with the Loch Droma date.

Three further sites have been investigated in Wester Ross in terms of a Lateglacial stratigraphy: Glasscnock (Robinson, 1977), Loch Clair (Pennington et al., 1977) and Loch Maree (H.H. Birks, 1973) but only the first of these has a complete sequence of Lateglacial sediments. At Glasscnock, zones A and B are clearly present (G1 and G2) with a *Rumex* zone followed by a zone of *Empetrum*. An interstadial oscillation in pollen spectra is also registered in this profile. The *Artemisia* zone C of the regional pollen zone stratigraphy is not, however, present. The increase in values of *Artemisia* pollen throughout much of the Scottish Highlands during the Loch Lomond Stadial has been noted above and there is a marked regional difference in representation of *Artemisia* during this period in the North West (Pennington et al., 1972; Macpherson, 1980; Tipping, 1985). Here pollen profiles show low percentages of *Artemisia* to the west of the regional watershed but increased representation to the east. Given that some species of *Artemisia* are chinophobous, the variations in its occurrence have been related to the extent of snow cover and hence amount of precipitation (see below), those areas with low *Artemisia* values such as Glasscnock (Robinson, 1977) being thought to be in areas of high precipitation. If this assertion is viable, then the increased representation of
Artemisia at Loch Droma (Kirk and Godwin, 1963) in comparison with Glasscnock implies a north-eastwards decline in precipitation across Wester Ross during the stadial. Zone C of Pennington et al., (1975) is an Artemisia zone but since this taxon is only really significant in the eastern sites the term is apparently less appropriate in the west.

Robinson (1977) equates the local pollen zone G4, at Glasscnock with the transition zone at Loch Sionascaig. At Loch Clair, the pattern of early vegetation development appears to be closely reflected where both A and B zones are present with Rumex followed by Empetrum constituting the dominant taxa including high Lycopodium, ferns grasses sedges birch and willow.

Robinson (1977) considered that the vegetational development at sites in Wester Ross are thought to have been representative of patterns established throughout the North West Highlands and that these were initially somewhat different from contemporaneous evolutionary trends on Skye but recent research has revoked this assertion. Walker et al., (1988) examined existing pollen profiles on Skye (c.f. Birks, 1973; Williams, 1977) and identified and analysed three new cores containing Lateglacial sediments. From this research they were able to produce two differing regional pollen assemblage zones for eastern Skye and South West Skye. Walker et al. (1988) point out that these new Lateglacial pollen diagrams show a marked overall similarity to those in Wester Ross (Pennington et al. 1972; Birks, 1984) to those further south in Argyll (Tipping, 1984) and especially to the pollen profile from Mull (Walker and Lowe, 1986).

There is a limited amount of information available concerning the early Postglacial development of vegetation in the north west highlands. Only two sites in Wester Ross have a full Flandrian pollen profile: one from Loch Maree (Birks, 1972) and another from Loch Clair (Pennington et al. 1972). Various sites at Beinn Eighe (Dumo and McVean, 1959, Loch Droma (Kirk and Godwin, 1963) and in the Torridon-Applecross area contain only part of the Flandrian sequence. In Sutherland a number of sites have also been recorded.

Where the earliest Flandrian sediments are present in Wester Ross, pollen taxa suggest that initial vegetation development there was dominated by an open dwarf shrub heath, the very earliest samples having significant values of Gramineae, Rumex, Huperzia selago, Salix herbaceae and Empetrum. Subsequent to this, a rapid increase occurs in values of Juniperus together with a decline in Empetrum. The rise in Juniperus values is attributed to either migration of the shrub into the area or to increased flowering (see above) of junipers already present as a response to climatic amelioration. The dates for the earliest vegetation changes are not firmly established but at Loch Maree and Loch Clair it is clear
that the early phase of juniper dominance ceased by c. 9 ka BP. Extrapolation of the linear sedimentation rate in the Loch Maree core to the base of the Flandrian indicates a lower age of 9.8 ka BP for these events and this is thought to indicate a relatively young age for the juniper peak compared with sites further south (Walker, 1984). After 9 ka BP there was a marked rise first in Betula and subsequently in Corylus/Myrica pollen values. High values of fern spores suggest that this woodland had an understorey of ferns. Pollen of Quercus, Ulmus, and Sorbus aucuparia together with spores of Pteridium aquilinum first appear or rise from low numbers in the Loch Maree core c 8.8 ka BP. During this period spores and leaves of Sphagnum Juncus seeds and Calluna vulgaris pollen are also recorded which would imply the local development of acidophilous mires in places. The duration of this phase of birch woodland is markedly different at Lochs Clair and Maree (despite the fact that they are only 5 km apart and 80 m different in altitude). At Loch Maree Pinus pollen values increased rapidly around 8.25 ka BP whereas such a rise did not occur at Loch Clair until 6.5 ka BP. This diachroneity was not apparent in the subsequent emergence of Alnus, which was well established by c. 6.5 ka BP at both sites (Birks, 1972; Pennington et al., 1972). The sequence of pollen assemblage zones at Loch Maree can be confidently equated with the NS regional pollen zones (Birks, H.H. 1973). Similarly, G4 at Glasscnock corresponds to NSI, although here Juniperus has no significant peak and G5 and G6 can be equated with NSII and NSIII.

The regional pattern of early Postglacial vegetation development in Sutherland is also reflected at sites in Sutherland. Moar (1969) for example defined zones equivalent to NS I-IV, although some minor differences were evident such as the early continuous presence of Alnus. These pollen assemblage zones are also present in Skye (Walker et al., 1989).

Vegetation development during the Lateglacial and Postglacial periods appears on the whole to have followed a consistent broad pattern throughout the mainland and inner islands of north-west Scotland. The Lateglacial is characterised by an initial period of herb-rich grass or sedge heath succeeded by the development of a treeless heath with widespread and abundant Empetrum. A minor climatic oscillation is implied near the beginning of the Lateglacial Interstadial by a brief decrease in pollen representation of woody plants and many herbs and an associated increase in the amount of Rumex. A vegetational revertance to an open vegetation cover dominated by ruderals, grasses and sedges subsequently occurred during the Lateglacial stadial. This was followed by a sequence similar to that of the beginning of the Lateglacial Interstadial with the development of a dwarf shrub heath and an expansion of juniper preceding the development of a tree cover and the establishment of a mixed deciduous woodland.
2.6 Conclusion

The Late Quaternary in the Scottish Highlands is characterised by marked environmental change. The nature and extent of glacial episodes during this time is well recorded for some areas but significant gaps in the record are apparent and a more robust chronology of such environmental changes is required for particular areas and for Scotland in general. Further integration of the geomorphological and biostratigraphical evidence may yield a more coherent picture of the nature and causes of Late Quaternary environmental change.
CHAPTER THREE

The study area

3.1. Introduction.
Until the present study, the area of southern Ross-shire stood out as one of the remoter parts of highland Scotland that had never been studied in any detail in terms of Quaternary history or glacial and periglacial geomorphology. This was perhaps primarily due to its remoteness and mountainous terrain. As described in chapter 1, the evident lack of information for this area meant that southern Ross-shire presented itself as an obvious site for further research in this field. The extent of the proposed field area was also self-determined to some extent as it was logical to fill in the gap between sites where the Quaternary history was more firmly established. The aims of this chapter are first, to show the location and extent of the study area, to outline its physical characteristics and geology, and to discuss the previous literature concerning its glacial and environmental history.

3.2. Location and topography.
The area studied comprises all of southern Ross-shire and includes some of the most north-westerly parts of western Inverness-shire. It is bounded by Glen Carron and Strath Carron in the north and its southern extent is represented by the mountains to the south of Glen Shiel, consisting of the Cluanie Ridge and The Saddle. The field area extends from Loch Carron and Loch Duich on the west coast to the lower parts of the eastern valleys that descend into Strathglass. It is defined approximately by longitudes 4°45' and 5°25' W and lies between the latitudes of 57°7' and 57°25'. The Great Glen Fault is less than 20 km to the south of the field area and Wester Ross, which encompasses Torridon and Applecross, is immediately to the north of it (Figure 3.1 in folder).

Almost all of the area consists of glacially dissected plateaux, which culminate in several mountain massifs and ridges that attain altitudes in excess of 1000 m. In addition to thirty five peaks over 1000 m, a further thirty are over 900 m. Mountains and high ridges dominate the landscape throughout. The highest summit is Carn Eighe (1183 m) which is one of several peaks (e.g. Sgurr nan Ceathreamhnan 1151 m; Mam Sodhail 1180 m) along a long ridge forming the Glen Affric Hills. This range is one of a series that trend west-east. To the south of this, the ridges of first the Five Sisters (Sgurr Fhuaran, 1068 m; Sgurr na Ciste Duibhe, 1027 m) and in the extreme south, the Cluanie Ridge, with peaks Aonach air Chrith (1021 m) and the Saddle (1110 m), also trend west-east. Towards the south-west, Ben Attow (1032 m) an isolated summit with an extensive plateau stands out as one of the most massive hills in the area and two further ridges
trending north-south occur in the south-east whose principal peaks include A' Chralaig (1120 m) and Sgurr nan Conbhairean (1110 m).

To the north of Carn Eighe, many of the mountains surrounding the Monar Basin are over 1000 m although in more general terms the height of the land decreases. Here, the more or less denuded, highland mountain plateau falls towards the north-east. The least dissected part of the plateau is the area surrounding the Monar basin. The Mullardoch Hills, to the south of the Monar Basin include An Riabhachan at 1129 m and Sgurr na Lapaich at 1150 m whilst the Strathfarrar hills to the north and north-east attain similar altitudes (e.g. Maoile Lundaih, 1104; Sgurr Fhuar-thuill, 1049 m; Sgurr a' Choire Ghlaigh 1083 m). In contrast to much of the central and southern parts of the field area dominated by numerous glacially entrenched troughs trending west-east across the study area with a general 'lofty and rugged' character (A. Geikie, 1893), the land towards the north-east and north-west becomes rather lower and adopts a more undulating lowland. Parts of this area consist of ice-scoured rock outcrops and small lochs or 'knock and lochan' topography with low rocky knolls and small lochs, with the bedrock in places mantled by peat and drift. The Sgurr a Mhuilin massif, three of whose peaks reach over 900 m form another isolated high level area in the north-east corner of the area.

Most of the principal valleys trend west-east and clearly assume the form of glacially eroded troughs. The watershed assumes a north-south position, and is located for example, at a height of 193 m in Glen Carron, 198 m at the head of the Monar Basin and at 273 m in Glen Shiel. The ground to the west is drained by the Carron, Ling, Elchaig and Shiel and that to the east by the upper parts of the Bran, Meig and Orrin (all tributaries of the Conon), whilst the region further south is drained by the Farrar, Cannich, Affric, Abhain Deabaig and the River Moriston. Many of the west-east aligned troughs have been overdeepened by glacier ice and the resultant rock basins are now occupied by lochs or fjords. In the far north of the field area, the Rivers Carron and Attadale reach the sea at Loch Carron, a formerly glaciated sea loch which constitutes a major fjord along this part of the coastline in this region. Similarly, the River Shiel enters Loch Duich in the SW, another, although slightly smaller fjord.

The coastline between the two fjords is essentially outwith the boundary of the field area. Several of the eastern valleys have been dammed to increase the area occupied by former lakes such as Lochs Monar, Mullardoch and Cluanie, and other lakes have merged due to damming. Somewhat smaller lochs are found occupying areas of morainic hummocks (e.g. An Gead Lochs in the upper Monar basin) and others occur in corries (e.g. Loch an Fraoch-chroie, NH 056 250; Loch Coire nan Dearcaig, NH 072 230).
A wide range of glacial deposits and landforms of glacial erosion affect the topography. Finely polished rock surfaces, corries with steep backwalls, evidence for small scale and major breaches and troughs all testify to the importance of ice in modifying and trimming the pre-existing landscape. Glacial, periglacial and fluvioglacial deposits mask the underling bedrock in other places. It would also appear that glaciation was instrumental in changing the patterns of drainage.

The topography of this area can be readily contrasted with that of areas of Wester Ross. Much of southern Ross-shire is composed of the metamorphic rocks of the Moine Series (see section 3.3.) which contrast with the comparatively unaltered Torridonian and Cambrian sediments of Torridon and Applecross to the north. Broad summit ridges or plateaux, and on the whole gentler slopes of the Moine schist area contrast with the impressively precipitous mountains to the north. The nature of the geology in the field area is considered in more detail below.

3.3. Solid Geology.

The field area lies within a region consisting primarily of crystalline metamorphic rocks in which the siliceous and pelitic schists of the Moine Series are dominant. It is located entirely to the east of the Moine Thrust zone, which separates ancient metamorphosed basement rock partly overlain by Precambrian and Cambrian sediments, from regionally metamorphosed sediments displaced westwards by at least 15 km (and possibly as much as 120 km) during the Caledonian Orogeny and representing the most prominent of a series of Late Caledonian overthrusts (Johnson, 1983). The Moine Thrust separates the Hebridean Craton to the west from the orthotectonic zone of the Caledonides in which the field area lies. This latter zone comprises mainly Moinian metasediments - the siliceous and pelitic schists of the Moine Series, but also inliers of altered Lewisian rocks. Granitic plutons and metamorphosed igneous rocks on the Moine and Dalradian Series are also present.

Metamorphosed sediments of the Moine Series crop out over a large area of the North West Highlands from the line of the Moine Thrust, eastwards and southwards to where in the southern highlands they dip beneath the Dalradian. In several parts of the North West Highlands including southern Ross-shire and western Inverness-shire the Moines are spatially associated with the rocks of the Lewisian type which are involved in the same fold movements (Figure 3.3. Johnson, in Craig). The rocks of the Moine Series are collectively known as Moine Schists. Their distribution is shown in Figure 3.2. For the most part these comprise psammitic-quartz-feldspar granulites, pelitic muscovite gneisses and mica-schists. The Moine schists often present a banded aspect owing to the alteration of more quartzo-felspathic rocks with more micaceous layers. This small-scale
banding reflects in miniature the composition of the Moine Series by alternating granulite and mica-schist groups.

As the early researchers noted, the intensity of the regional metamorphism in the North West Highlands increases rapidly as the Moines are traced eastwards from the Moine Thrust outcrop. Near the thrust the pelitic members of the Moines are fine-grained biotite or garnet-bearing schists, but within a few kilometres, these give place to sillimanite-bearing migmatite gneisses.

Figure 3.2
Solid Geology of southern Ross-shire
Although there has now been more than eighty years research in the subject, neither the stratigraphic age nor the age of the succession of the Moine lithostratigraphic units has been established. There are however grounds for believing that the Moine Series can be subdivided chronologically into two groups: the 'old' Moines which appear to have been originally deposited c. 1300 - 1200 Ma and which were subject to alteration and deformation during both the Grenville and Caledonian orogenies; and the 'young' Moines, which may be roughly equivalent to the Torridonian group to the north in age of deposition which experienced alteration only during the Caledonian orogeny.

The status of the Moine rocks has also proved difficult to establish. Whereas it is now clear that the Moines postdate the Lewisian rocks which were reworked with them and pre-date the Dalradian, their time of deposition within the 1000 year interval between the Laxfordian (c. 1700 Ma) and the Dalradian (c. 700 Ma) is uncertain (Harris, 1983).

The Lewisian gneiss consists of highly deformed and metamorphosed rocks that form a stable crystalline basement. Scourian and Laxfordian rocks modified by Caledonian deformation and metamorphism (c. 1000 - 400 Ma) are important to the west of the Moine Thrust but are also found to the east of the thrust where they form inliers in the basement of the Caledonian orogenic belt. Both the Scourian and Laxfordian rocks are mostly banded gneisses that have been metamorphosed and migmatised at high temperatures and pressures. An increase in the stage of metamorphism of the Lewisian gneiss is met with in passing eastwards from the Moine Thrust over successive inliers.

The Lewisian basement lies unconformably below the Moine Series and is seen within the Caledonides of the Northern Highlands as a number of autochthonous or allochthonous masses ranging from a few hundred metres to 50 km in length. It is clear that prior to the deposition of the Moine sediments they formed a gneiss complex lithologically and geometrically quite like the Scourian complex of the Hebridean craton and penetrated, in some areas by basic dykes. The dominance of Na over K and rather low levels of incompatible elements serve on the one hand to distinguish basement outcrops from the Moine metasediments (Winchester and Lambert, 1970; Winchester, 1971; Harrison and Moorhouse, 1976; Johnstone et al, 1975). An Rb-Sr isochron of c. 2700 Ma for an outcrop at Scardroy (Moorbath and Taylor, 1975) and a number of mineral ages up to 2200 Ma from Glenelg confirm the dating of the basement as Scourian (Moorbath, cited by Watson, 1975).

In central and southern Ross-shire the Lewisian rocks form a sheet-like mass separating the Morar division from the Glenfinnan division. The various Lewisian outcrops are portions of the sheet detached by deformation and erosion. Sutton and Watson (1962)
have pointed out that the Lewisian rocks at Scardroy rest upon an uninverted Moine sequence (Lower Morar Psammite) some 2 km thick. After Sutton and Watson (1962), the rocks in southern Ross-shire have been regarded as a large thrust sheet. Here, 'inliers' of complex gneisses of Lewisian type are most widely developed towards the east where the largest mass is found in the basin of the Meig and forms part of the Sgurr a Ghlaist Leathaid. Other small masses occur in Glen Orrin, Strathfarrar, north of Sgurr na Lapaich and around the western end of Loch Monar.

The status of the Lewisian inliers, like the Moines, has however, long been the subject of controversy. Originally they were interpreted by Peach as stratigraphical Lewisian inliers in the Moines, but later their pre-Moine age was disputed notably by Reed (1934). They are now regarded as Lewisian rocks that form either thrust slices or isoclinal fold cores. The proof that they are Lewisian has been demonstrated at Glenelg and between Loch Carron and Loch Houm. The rocks of both groups are over almost all the area granulitic and possess common foliation which has in most areas obliterated original structures.

Intruded into the rocks of the Moine group are granitic plutons associated with the Caledonian orogeny. In the southernmost part of the field area lie two such granitic intrusions, those of the Cluanie Granodiorite and Glenelg-Ratagain complex which constitute New granites, introduced after regional metamorphism and folding. The Cluanie pluton consists almost entirely of granodiorite and at the Glenelg-ratagain complex there is an intrusive sequence of granite, diorite, syenite and adamellite. These intrusions have caused considerable deformation of their country rocks (Leedal, 1952; Nichols, 1956).

Other outcrops of igneous rocks can be found in southern Ross-shire in places. Narrow dyke intrusions are common in places, for example in Glen Strathfarrar, Glen Orrin and Glen Lichd and especially in the region of An Riabhachan and Sgurr na Lapaich. These intrusions are typically amphibolite and hornblende schists in the east, epidiorite occurring in narrow bands is also present and porphyrite, lamprophyre and rhyolite are particularly prevalent towards the south of the study area. Granitic gneiss of the Moine Series also occurs, the largest outcrop of which is located in the south west of the field area in Glen Moriston.

The Devonian sediments found to the north in Wester Ross do not enter the field area and there is relatively little lithological differentiation here.

With regard to the structural aspects of the geology of southern Ross-shire, several of the largest structural features are due, directly or indirectly, to dislocation. Of these, the
Strathconan Fault is by far the most important. It enters the region at the foot of Gleann Meinich and, following a direct south-westerly direction, continues down to just south of the head of Loch Duich. This has clearly been a powerful line of disruption: it is responsible for the east to north-east trend of the valley of the Meig below Inverchoran and also the further alignment of Gleann Chaorain, Glen Orrin above Loch na Caoidhe and the glen that falls from the head of the Orrin to Luib an Inbhir in the Monar basin. A series of escarpments trending south-west from Altfearn and along the valley of the Allt Riabhachan indicates the further course of the fault to the south of Loch Monar.

A fault occurring as the basin of Loch Beannacharain in the Meig valley coincides with the eastward extension of a large fault that has determined the northern shore of Loch Maree in Wester Ross. The western part of Strathcarron from Lair to the head of Loch Carron and the Eas Ban have also been identified as a fault-lines. Towards the south, a north-west to south-east trending fault is present to the east of the Five Sisters in Glen Shiel.

The mountains of the Moine Schist area present, as a rule, and especially where they are composed of the more siliceous members of the series, broad summit ridges or plateaux, while smooth and grassy slopes separate the craggy rather than precipitous glens and corries. In those places where the inliers of Lewisian gneiss penetrate the landscape, the surface features are typically of undulating rocky plateau, interrupted by abrupt craggy hills and deep lochan-filled hollows, for example at Scardroy and in Glen Orrin. Where the course pelitic gneiss occurs at high mountain ground and the erosion of the heads of the corries has extended far back into the summit ridge, narrow shattered crests and deep rock-walled corries have formed such as those on an Riabhachan and Sgurr na Lapaich and some of the hills of the West Monar Forest.

Much of the field area is covered by glacial and periglacial sediments, deposited or formed during the last and previous glacial phases. Past research pertaining to these is examined in the following section.

3.4. Glaciation in the study area.
3.4.1. Introduction.

The earliest published reference to glacial phenomena in the study area appears in the third edition of Archibald Geikie's Scenery of Scotland (1901). Geikie described (p. 294) a group of 'marginal terraces' at Achnasheen in the northernmost part of the area, and attributed these to formation at the margin of a lake dammed by a lobe of ice that emanated from the Fannich mountains to the north. He also inferred (p. 259) that glacier
ice in the Great Glen to the east of the study area had been fed by 'large glaciers' that occupied Glen Moriston and Glen Urquhart, but furnished no supporting evidence for the former existence of these glaciers.

The first focused accounts of glaciation in Ross-shire are to be found in the Memoirs of the Geological Survey, written around the turn of the century by B.N. Peach and his co-workers (1910, 1913a, 1913b, 1914). The majority of the study area is covered by the memoir for Central Ross (Peach et al., 1913a). The easternmost part of the area is described in the memoir for Beauly and Inverness (Peach et al., 1914) and a slightly earlier publication which covers Glen Elg, Loch Alsh, and the south-east part of Skye deals with the southernmost part of the study area (Peach et al., 1910). In these memoirs Peach et al. (1910, 1913a, 1913b, 1914) described various Quaternary landforms and deposits and proposed a sequence of events exemplified by specific field evidence within the area. The earliest stage is described as a period of maximum glaciation when glacier ice probably covered the whole area. The second phase was envisaged as one of 'confluent glaciers' when extensive glaciers extended westwards and eastwards from the line of the present watershed with the higher mountains rising above the ice as nunataks. They envisaged the final stage as one of limited corrie and valley glaciation. The evidence for these stages is considered in turn below.

3.4.2. (Late Devensian) Ice Sheet Glaciation

Published information concerning the last (Late Devensian) ice sheet in Southern Ross-shire is scant. Peach et al. (1913a) noted the lack of striae in this area and suggested that striae relating to ice sheet glaciation have been subsequently obliterated by the action of frost in Lateglacial and postglacial times. A few striae indicating the direction of former ice movement during the period of maximum glaciation are mentioned, however. In the north of the field area striae were found at the summit of Sgurr na Feartaig (862 m; NH 055 454) and striae oriented in a north-easterly direction were identified to the north of Creag na h-Iolaire (578 m; NH 175 498) and to the east of Cam Gorm (875 m; NH 135 500). Farther south, and immediately to the north of Loch Alsh, striae recording former ice movement of ice 'a few degrees to the north of West' were similarly assigned to the period of maximum glaciation. With regard to glacial deposits, Peach et al. (1913a) identified distinct tills beneath fluvioglacial deposits in Glens Attadale and Ling. These tills they inferred to be associated with the ice sheet glaciation, but no differentiation was made between deposits of different ages.

Subsequent accounts make little mention of evidence for ice-sheet glaciation in southern Ross-shire. Sissons (1967a) implied that the whole of the area was covered by glacier ice during maximum ice sheet glaciation but made no specific reference to southern Ross-
shire. Peacock (1975) made cursory references to Late Devensian ice sheet glaciation in the area, particularly regarding former ice movement directions in Glen Affric and Glen Moriston. He recorded eastward-directed striae of ice-sheet age on high ground north of Loch Affric, and distinguished these from later striae that he presumed to have been cut during the Loch Lomond Readvance. Peacock also suggested that the Carn Eighe massif was the centre of vigorous glaciation during the last ice sheet maximum but provided no supporting evidence. More recently, in a study of reconstructed Pleistocene ice-sheet temperatures, Gordon (1979) considered the distribution of areal scouring across southern Ross-shire. He attempted to demonstrate a relationship between spatial differences in reconstructed basal ice-sheet temperatures and the distribution of areal scouring by comparing the distribution of knock and lochan topography for the southern part of the field area with predicted basal ice temperatures. The results, however, proved inconclusive, partly on account of Gordon's assumption of ice-sheet margins much more extensive than those now known to relate to the limits of the last (Late Devensian) ice sheet (cf. Sutherland, 1984; Bowen et al., 1986).

3.4.3. Confluent glacier stage.
The wastage of the last ice sheet in southern Ross-shire from its maximum is described by Peach et al. (1913a, 1914) in their accounts of a 'confluent glacier' stage. Peach et al. provided little concrete evidence pertaining to this stage in relation to the field area. They suggested, however, that at this time the Monar Basin was a major centre of ice dispersal from which several outlet glaciers had radiated, spreading out to form a continuous glacier cover over lower plateau areas. Evidence for such a cover was inferred from the 'ice worn summits' of Meallan Buidhe (555 m; NH 037 377), Meallan Odhar (570 m; NH 053 384), Beinn Dubh (591 m; NH 037 377) and other nearby hills of lower elevation that support glacially scoured bedrock. From such evidence, they adduced an eastward movement of confluent glacier ice into Glen Strathfarrar. This evidence, however, does not necessarily reflect a separate stage and may merely indicate that the Late Devensian ice sheet attained at least the altitude of these hills. Apart from these references to the Monar region, evidence for a 'confluent glacier' stage was not detailed for other parts of the field area. Although there has been no subsequent mention of evidence relating to this stage, Charlesworth's (1955) wide-ranging monograph covering the Late Glacial history of the Highlands and Islands of Scotland described the deglaciation phases of 25 named glaciers (and numerous tributary glaciers) in southern Ross from their supposed ice-sheet maximum position. Charlesworth did not, however, cite any form of evidence and the detailed maps depicting the reconstructed glaciers can only be considered speculative.
No evidence has been provided for a discrete and widespread readvance in the field area occurring between the last ice sheet maximum and the readvance of local glaciers during the Loch Lomond Stadial, despite the fact that evidence has been found in the area north-west of the study area for an intervening readvance (the Wester Ross Readvance) that interrupted ice-sheet deglaciation (Robinson and Ballantyne, 1979).

3.4.4. The Loch Lomond Readvance.

Although Peach et al. (1913a) outlined a threefold subdivision of glacial events, they did not always make a clear distinction between the second 'confluent glacier' stage and the final 'valley and corrie' glacier phase. Features that were assigned merely to a period of advanced ice-sheet wastage included terraced lateral moraines around Meall an Flùichard (NH 067 496), and along the eastern slopes between Beinn Dronaig (NH 037 382) and Beinn Tharsuinn (NH 056 433). Peach et al. also described moraines at the head of Loch Carron and at Achnasheen. These they attributed to a later phase of the 'confluent glacier' stage, and they envisaged that remaining ice retreated up the glens leaving successive lines of moraines until only small glaciers remained in the corries. In the Monar region, however, their 'valley glacier' stage is depicted as one of relatively small, locally-nourished glaciers, six of which were believed to have existed here (Figure 3.3). The evidence in support of these was given as striae, hummocky moraines, boulder trails and spreads of drift.

**Figure 3.3**

Distribution of 'valley glaciers' in the Monar Basin (after Geological Survey (1913a))
Research carried out over the past 30 years makes it possible to consider the early work of Peach and his co-workers in terms of a broader chronological framework that has been developed for the Scottish Highlands as a whole. During the 1960s, evidence was found that indicated that the Loch Lomond Readvance, first identified by Simpson (1933), had covered large parts of the Scottish Highlands. Sissons (1967) refers to this last period of glaciation in 'The Evolution of Scotland's Scenery' but in relation to the field area it is mentioned only briefly: 'in the western mountains from southern Ross to Loch Lomond the altitude coupled with heavy precipitation produced a greater accumulation of ice than elsewhere and the main valleys were occupied by large glaciers fed by many tributary glaciers'. This statement was, however, unsupported by field evidence. Subsequent research has nevertheless provided some support for Sissons' estimation of the extent and nature of the Loch Lomond Readvance in southern Ross-shire. Peacock (1975), for example, has argued that former ice limits in the Glen Moriston-Glen Affric area are of Loch Lomond Readvance age, on the grounds that '...the geomorphological framework is similar to that known for Loch Lomond Readvance glaciers elsewhere in the Highlands...' and that '...the district is contiguous with ground to the south for which an intensive glaciation of this age has already been inferred (Peacock, 1970)'. Neither argument can be considered conclusive, however, and Peacock offered no direct evidence for the age of the ice limits he identified in this area.

Peacock proposed that striae in Glen Moriston indicated the former presence of an eastward-flowing glacier of Loch Lomond Readvance age, and placed the limit of this glacier 3-4 km east of Dundreggan on the evidence of 'a well-marked end moraine on the south side of the valley' (Figure 3.4). He inferred from the presence of high-level striae and the transport of erratics of Cluanie Granodiorite that ice at this time had passed north-eastwards over a 730m ridge into Glen Doe, which he believed to have contained a glacier that was confluent with that in Glen Moriston. He believed that the contemporaneous ice shed at the head of Glen Moriston was located 'about 1 km east of the present principal watershed' but furnished no evidence in support of this claim.

Similarly, Peacock placed the estimated Loch Lomond Readvance ice-shed some 4-5 km east of the present watershed at the head of Glen Affric (Figure 3.5), but noted that 'its precise position has yet to be determined'. He was unable to pinpoint the eastern limit of the Glen Affric glacier, but from the evidence of striae he adduced that glacier ice had swept eastwards over a col at 610m altitude into the valley of the Abhainn Deabhag, terminating near the mouth of that valley near Tomich.
Figure 3.4.
Distribution of Loch Lomond Readvance glaciers in the G. Moriston area (Sissons, 1977)

Figure 3.5
Former ice limits in the Glen Affric area
In this context, it is useful to consider now research relating to the extent of the Loch Lomond Readvance in northern Ross-shire. To the north-west of the study area, Robinson (1977, 1987) mapped the limits of the Loch Lomond Readvance in Applecross and the mountains to the south of Glen Torridon. In the former area she identified seven corrie or valley glaciers of limited extent, in the latter a complex icefield with numerous outlet glaciers. She also provided strong stratigraphic evidence in favour of a Loch Lomond Stadial age for this icefield complex. Cores taken from a peat-filled kettle hole at Glasscnock (NG 866 461), 2 km outside a moraine that marks the limit of one of the outlet glaciers, contained organic sediments relating to the Lateglacial Interstadial. Cores taken from a similar basin immediately inside the same moraine limit at Druim Dubh (NG 885 471) revealed only Flandrian sediments, which implies that the latter site (but not the former) was covered by glacier ice during the Loch Lomond Stadial.

The study area under investigation here terminates to the north-west in Strathcarron, which also marked the south-eastern limit to the area mapped by Robinson. In lower Strathcarron she mapped a long, arcuate drift ridge, which she interpreted as the terminus of a Loch Lomond Readvance glacier that flowed into Strathcarron from the icefield to the north. Her mapped glacier limits also depict a second lobe of ice debouching into Strathcarron from the valley of the River Lair. From the evidence of various fluvioglacial features indicative of ice stagnation, Robinson deduced that these two outlet lobes had coalesced in Strathcarron, and that the limit of this 'Strathcarron Glacier' was marked by a belt of lateral moraines on the north side of the valley. Robinson envisaged that the equivalent glacier margin on the south side of Strathcarron descended south-westwards along the lower northern slopes of Carn Mor (NG 971 432), but reported no field evidence in favour of this interpretation. Similarly, although her map depicts ice from corries north of Sgurr na Feartaig (805m; NH 038 450) feeding northwards into the 'Strathcarron Glacier', she provided no field evidence for this, nor, indeed, for her proposed upvalley limit to the 'Strathcarron Glacier' near Craig (NH 139 493). Her account therefore leaves unanswered the question of whether or not outlet glaciers nourished in southern Ross-shire fed glacier ice in Strathcarron during the Loch Lomond Stadial (Figure 3.6).

At the north-eastern end of Glen Carron (the upvalley continuation of Strathcarron) Sissons (1982b) undertook detailed mapping of the outwash deltas, lake shorelines and former ice limits at Achnasheen (Figure 3.7), features previously mentioned and interpreted in terms of the former existence of an ice-dammed lake by Geikie (1901) and Peach et al. (1913b). He inferred that a thick ramp of drift across Strath Bran marks the westward limit of a tongue of ice that extended southwards from the Loch Fannich basin and impounded an ice-dammed lake in upper Strath Bran. Sissons also mapped evidence in the form of a lateral moraine, drift mounds and spreads of boulders, all of which
Figure 3.7
Former ice limits around Achnasheen

1. Drift mounds
2. Delta terrace
3. Large kame terraces
4. Kettles
5. Lake shoreline
6. Cross valley moraines
7. Meltwater channels
8. Contours
9. Glacial limit
suggest that ice tongues from the Loch a' Chroisg valley to the west and Glen Carron to the south-west terminated in this lake, such that the outwash from these glaciers formed impressive ice-marginal deltas (Benn, 1989). That the Strath Bran lobe represented a glacial readvance was demonstrated by the presence of lake sediments underlying glacigenic deposits at the margin of the drift ramp.

Sissons considered that the dimensions of the three ice masses described above are inconsistent with those of glaciers formed during the Loch Lomond Stadial, and suggested that the 'Achnasheen Readvance' may correlate with the Wester Ross Readvance, though this would seem to imply a very asymmetric icefield at the time of the latter (Ballantyne et al. 1987). Moreover, Sissons was unable to discount the possibility of a Loch Lomond Readvance age for the glaciers that terminated in the ice-dammed lake at Achnasheen. He himself was unable to find evidence of pre-Loch Lomond Stadial sediments by coring through peat inside the limits of the Strath Bran lobe, and Pennington et al. (1972; also Pennington, 1977) failed to find organic Lateglacial sediments in Loch a' Chroisg, which lies immediately inside one of the Achnasheen moraine limits. It thus remains possible, as suggested by Ballantyne et al. (1987, p 26-27) that the former ice limits at Achnasheen represent the extent of the Loch Lomond Readvance in this area. If so, one implication is that glacier ice may have occupied all of Glen Carron and Strath Carron at this time, terminating to the south-west near Loch Carron at the end moraine identified by Robinson (1977, 1987) and to the north-east at the former ice limit identified by Sissons (1982b) near Achnasheen.

Previous research regarding the extent of the Loch Lomond Readvance is thus restricted to the periphery of the study area; apart from the early work of Peach et al. (1913a) in the Monar basin no attempt has been made to identify the limits of locally-nourished glaciers in the central part of southern Ross-shire. Moreover, most of the research accomplished to date lacks an adequate stratigraphic or chronological framework. As a result it is not as yet certain whether this area supported one or more glacial readvances of the last Scottish ice sheet from its readvance maximum extent.

The above review suggests four research priorities with regard to the glacial history of southern Ross-shire. First, the directions of ice movement and vertical extent of the last ice sheet in the area require to be established. In the light of recent findings by Reed (1988), which suggest that many mountains in northern Ross-shire remained above the surface of the last ice sheet, it is of particular interest to learn whether ice-sheet nunataks existed in southern Ross-shire also. Second, it remains to be established whether the area contains evidence for glacial readvances that interrupted ice-sheet retreat, as Sissons (1982b) has proposed for the readvance limits at Achnasheen. Third, the true extent of
the Loch Lomond Readvance in this area remains to be determined. Sissons' (1983) summary map of the extent of the Loch Lomond Readvance in Scotland depicts southern Ross-shire as supporting an extensive icefield and associated outlet glaciers (Figure 1.1) but appears to be based on the limited and contentious information available from the research on the extent of the readvance on the periphery of the area, and is thus largely speculative. Finally, within the vast area of southern Ross-shire there is no direct stratigraphic or biostratigraphic evidence to provide a basis for establishing a chronology of glacial events. It is towards fulfilment of these four research priorities that much of the research reported in this thesis is directed.

3.5. Holocene landforms and environment.

Following the retreat of Loch Lomond Readvance glacier ice from southern Ross-shire, a variety of processes have operated to modify the glacial landscape. Geomorphologically, the most important of these have been slope processes and particularly the occurrence of landslides and the formation of alluvial features.

Landslides have occurred throughout the field area but are concentrated in a number of areas such as the Glenshiel hills, the Carn Eighe massif, the Mullardoch Hills north of Glen Cannich and in the valley of the River Meig. Evidence for landslides here shows that they were concentrated in the vicinity of the Strathconon fault, a major structural feature, that runs north-east to south-west through southern Ross-shire. Several landslides, tens of kilometres long are present. Such features exist in Glen Lichd, on the south side of Ben Attow, on the south-facing slope of An Riabhachan and on the east-facing side of Mullach na Deirigan. Other features include alluvial fans (e.g. in Glen Elchaig), talus sheets and present day periglacial features such as turf-banked terraces and ploughing boulders are clearly forming. Peat is locally extensive in valleys and on plateaux throughout the field area and is frequently hagged.
CHAPTER FOUR

Methodology: the mapping of the glacial and periglacial evidence

4.1 Introduction.
The aims of this chapter are threefold: first, to outline the procedures employed in mapping the field area; second, to describe the main forms of geomorphological evidence that have been used to reconstruct former ice limits and directions of ice-flow; and third, to discuss some of the problems associated with the mapping of such evidence and its use in reconstructing patterns of glaciation.

An extensive array of morphological evidence has been used to reconstruct former glacial limits and ice surfaces in upland Scotland (e.g. Thompson, 1972; Gray, 1972, 1982a; Gray and Brooks, 1972; Sissons, 1972, 1974a, 1977a, 1977b, 1979a; Sissons and Grant, 1972; Robinson, 1977; Ballantyne and Wain-Hobson, 1980; Cornish, 1981; Thorp, 1981b, 1984, 1986; Ballantyne, 1989a). In previous studies such as the aforementioned, however, the importance of each of the various forms of evidence has been given differing emphasis and the types of evidence used have clearly reflected the size and form of the ice mass being reconstructed. The main forms of glacial evidence that have been used to delimit former ice masses in Scotland include end, lateral, medial, fluted and hummocky moraines, erratics, striae and fluvioglacial landforms, which include meltwater channels and outwash sands and gravels. Indications of the direction of former ice flow are provided by erratics, striae, ice-moulded bedrock, roches moutonnées, streamlined and fluted drift and subglacial meltwater channels.

Relict periglacial landforms have also been employed to complement this evidence. In this context, the most important features include relict solifluction sheets, lobes and terraces, frost-shattered bedrock, fossil screes, blockfields and large-scale sorted patterned ground. The use of such evidence has been extended through the identification of periglacial trimlines, which provide an important additional means of delimiting the upper surfaces of former ice masses. Various studies have stressed, however, that it is the total assemblage of glacial, fluvioglacial and periglacial evidence that is significant in providing information regarding both the lateral and vertical extent of the ice (e.g. Thorp, 1986; Ballantyne, 1989a).

Geomorphological evidence relating to both the Late Devensian ice sheet and the Loch Lomond Readvance has been mapped in Scotland. With regard to delimiting Loch Lomond Readvance glaciers, on the evidence described, ice limits are most clearly defined
by end and lateral moraines. Such evidence is apparently most useful for delimiting small corrie and valley glaciers, but is often only significant in relation to determining their downvalley extent. Recent mapping has shown that the use of periglacial trimlines is valuable for establishing the thickness of former glaciers. This form of evidence has proved to be particularly useful in the mapping of large icefields or transection glacier complexes, and is therefore especially applicable to the present study. Furthermore, the evidence afforded by periglacial trimlines at high altitudes in the Scottish Highlands has recently been employed in attempts to delimit the vertical extent of the Late Devensian ice sheet (Reed, 1988; Ballantyne, 1990). Evidence for determining former ice-flow directions is present for both the Late Devensian ice sheet and for subsequent glacial advances or readvances. Sections 4.3, 4.4 and 4.5 below detail the methodological considerations associated with the mapping of relevant glacial and periglacial features.

4.2 Mapping procedures.

Vertical aerial photographs covering the entire field area at an approximate scale of 1:25,000 were examined before commencement of fieldwork with a view to identifying major landforms such as moraines, limits of thick drift and areas of periglacial frost-shattered detritus. Field mapping was subsequently carried out using 1:10,000 or 1:25,000 Ordnance Survey contoured base maps, depending upon the nature of the terrain to be mapped.

Much previous geomorphological mapping in Scotland has tended to be highly selective, focusing on major and unequivocal landforms such as end and lateral moraines. In view of the importance of less 'obvious' forms of evidence such as trimlines, it was considered necessary for this study to map the entire field area, both outside and within the area of proposed readvance limits. A greater range of symbols was therefore required than has been employed in the majority of earlier studies; Holocene and other features were noted in addition to landforms and deposits associated with Devensian glaciation and periglaciation. The symbols used are shown in Figure 4.1. The evidence recorded on the Ordnance Survey maps included types of drift, evidence for former directions of glacier movement (erratics, striae, roches moutonnees, ice-moulded bedrock, fluted and streamlined drift), evidence for former glacier limits (end and lateral moraines, the limits of drift against drift-free terrain, limits to hummocky drift and periglacial trimlines marking the upslope transition between ice-moulded and frost-shattered bedrock), relict periglacial features indicative of areas that remained above the limits of the last local glaciers and Holocene features that have developed since the final disappearance of glacier ice. Most of the symbols were conceived before the onset of fieldwork and these were supplemented by others established during mapping to cover the entire range of landforms that were encountered.
Figure 4.1
The symbols used in the geomorphological mapping

GLACIAL FEATURES
- Drift features
  - hummocky drift
  - area of mosaic hummocks
  - undulating/dissected drift (thick/thin)
  - moraine ridge
  - streamlined drift
  - drift limit
  - indistinct drift limit
  - fluted drift
  - erratics/glacially transported boulders
  - till sheet

Fluvioglacial features
- kame and kettle topography
- glacial outwash
- esker
- roche moutonnée
- meltwater channel

Bedrock features
- ice moulded bedrock
- ice moulded bedrock (directional)
- frost shattered bedrock
- frost shattered/moulded bedrock

TOPOGRAPHIC FEATURES
- boundary to a deposit
- edge of mapped area

OTHER FEATURES
- landslide and scar
- arete
- debris flow/cone
- raised beach deposits
- alluvium
- alluvial fan
- fault line
- fault
- site for comment in text
- free face
- gully incised in bedrock
- gully incised in drift

Periglacial features
- distinct periglacial trimline
- indistinct periglacial trimline
- deep frost shattered detritus
- incipient frost shattered detritus
- mass moved frost shattered detritus
- mass moved incipient detritus
- debris surface
- blockfield
- block slope
- talus
- relict talus
- solifluxion sheets
- sorted patterned ground
- relict active
- solifluxion lobes relict/active
- turf banked terraces
- boulder lobes
- solifluxed drift
The geomorphological mapping of the field area proceeded by working from a number of starting points: first by field checking and re-mapping of features previously identified on aerial photographs; second, from a search for relevant evidence cited by the Geological Survey within the field area; and third, by extending mapping from information provided by earlier studies on the periphery of the field area (e.g. Robinson, 1977; Peacock, 1970a; Sissons, 1977d). In many previous studies of glacier limits in Scotland, mapping has usually proceeded from the former ablation zones back into the former accumulation zones. In southern Ross-shire, however, this proved possible only in some areas because of the lack of end moraines or other unequivocal evidence for former glacier limits on low ground. It was therefore necessary in some parts of the study area to adopt a different strategy that involved first the identification of periglacial trimlines in former ice accumulation areas and subsequent extension of the mapping downvalley into the former ablation zones.

Aerial photographs were used simultaneously and in close conjunction with field mapping. This was primarily to check the location and altitude of landforms mapped in the field. Finally, when mapping was complete, the aerial photographs were employed to ensure planimetric accuracy in the transfer of field information to smaller scale Ordnance Survey maps.

The paucity of clear end and lateral moraines encountered in the field area meant that the reconstruction of former glaciers was particularly dependent on mapping of the whole suite of landforms and deposits present. In addition to the general traversing of the field area, particular procedures were adopted with regard to mapping certain features. In order to establish evidence for trimlines, for example, spurs were ascended or descended and changes in the characteristics of the bedrock and periglacial deposits were noted (cf. Thorp, 1981, 1986). Similarly, cols were traversed to establish whether they have been breached by ice during the Loch Lomond Readvance, by establishing whether these showed signs of unmodified ice-moulding or frost-shattering.

Exposures of superficial deposits were also examined wherever possible. Important differences in clast shape and lithology were noted and maximum thicknesses of till and fluvioglacial deposits were measured or estimated. Surfaces of exposed bedrock were examined for microscale features (e.g. striae and friction cracks). Where striae were located their orientation was determined using a compass. The coastal section of the area was mapped at 1:10,000 in order to identify evidence for former sea-levels and to ascertain their size and form for purposes of surveying.
Two factors inhibited the geomorphological investigation of the field area. Afforestation was the first problem. Many of the eastern valleys have been afforested since the introduction of hydroelectric schemes to the area. This was only partly compensated for by the availability of aerial photographs taken prior to planting. Second, there is extremely widespread peat throughout the area, although this is mainly restricted to ground below 500 m.

4.3 Glacial and fluvioglacial evidence.

4.3.1 Evidence for delimiting former glaciers.

4.3.1.1 Moraines.
Perhaps the most unequivocal forms of evidence used in the reconstruction of former ice marginal positions are terminal and lateral moraines (Plate 4.1). Such features have been mapped throughout the Scottish Highlands and many have been used to delimit the former maximal extent of Loch Lomond Readvance glaciers, for example in the North West Highlands (Sissons, 1977c, 1979a; Robinson, 1977), on Skye (Ballantyne, 1989a), Rhum (Ballantyne and Wain-Hobson, 1980), Mull (Gray and Brooks, 1972) and in the western Grampians (Peacock, 1971a; Gray, 1972, 1975 Thorp, 1984, 1986). Within the limits defined by end and lateral moraines linear morainic ridges may occur, that reflect recessional stages subsequent to the Loch Lomond Readvance maximum as the ice margin became temporarily stabilised during deglaciation. These were mapped with a view to establishing the extent of any 'stillstands' during deglaciation and correlating such events between valleys.

Such evidence is present only locally in the field area. The lateral limits of large Loch Lomond Readvance valley glaciers are marked by terminal moraines at Tomich as well as in Glen Moriston and Strathcarron. Corrie glacier moraines also occur to the north-west of Sguman Chòinntich in the west of the field area. Moraines therefore delimit only a small proportion of the proposed Loch Lomond Readvance extent. The absence of end and lateral moraines in many valleys may be attributable to lack of debris supply or erosion by meltwater or may be due to the glaciers not maintaining a steady state position for a sufficient length of time.

End and lateral moraines associated with readvances of the last ice sheet have been identified on Skye (Ballantyne, 1988) and Wester Ross (Robinson and Ballantyne, 1979) but equivalent forms were not identified in southern Ross-shire probably because this area is too close to the centre of former ice-sheet accumulation, and because the maximal limit of the later and less extensive Loch Lomond Readvance lies close to the west coast and to the eastern boundary of the field area.
Plate 4.1
Coire Mhuilidh lateral moraine, Strathfarrar

Plate 4.2
Hummocky moraines on a col at NH 159463
4.3.1.2 Hummocky moraines.
The term 'hummocky moraines' (or hummocky drift) has been widely used to describe moundy topography comprising steep-sided deposits, often strewn with numerous boulders. Hummocky drift is widespread in the Scottish Highlands and occurs throughout the field area. Such forms have been used extensively to delimit Loch Lomond Readvance glaciers in Scotland. This approach was first used by Sissons (1967a), who argued that the distribution of hummocky drift is often coincident with clear lateral and end moraines of probable Loch Lomond Stadial age. He therefore proposed that where end moraines are absent and where there is a clear hummocky moraine limit, such a limit could be used to infer the (maximum) extent of the Loch Lomond Readvance (Sissons, 1967a, 1973, 1974). This method has been subsequently employed by numerous other researchers in this context (e.g. Thompson, 1972; Gray and Brooks, 1972; Sissons et al., 1973; Sissons, 1974a, 1977a, 1977b, 1977c, 1979a, 1979b; Sissons and Grant, 1972; Robinson, 1977; Young, 1978; Ballantyne and Wain-Hobson, 1980, Cornish, 1981; Gray, 1982a; Thorp, 1984, 1986; Ballantyne, 1989a) but there has been much debate as to the validity of the use of hummocky drift alone to delimit former glacier margins (e.g. Sissons, 1972; Sugden, 1972; Clapperton et al., 1975; Clapperton and Sugden, 1977).

Two problems have emerged: first it has been argued that the distribution of hummocky moraines is not related to any particular readvance stage, but is dependent on the distribution of topographical locations conducive to the in situ stagnation of ice that had become isolated from active parts of a wasting ice mass (Clapperton and Sugden, 1977). Second, areas of hummocky drift have been mapped in some areas outside the Loch Lomond Readvance limits as defined by end moraines and associated landform assemblages (e.g. Sissons, 1977c). The absence of hummocky moraines in any area may simply reflect steep valley sides, a lack of debris or erosion by meltwater. However, the use of hummocky moraines to delimit former glaciers has continued, and such forms are still employed, albeit more cautiously, as a valuable indicator of the minimum extent of former local glaciers. There is, however, growing concern as to the significance of the distribution of hummocky drift and its mode of origin, and it has been used increasingly in conjunction with other forms of evidence: the presence of hummocky drift alone is no longer accepted as evidence of occupation by glacier ice during the Loch Lomond Readvance.

The genesis of hummocky drift is crucial to understanding of the nature and significance of these deposits. Sissons (1973, 1979b) proposed that hummocky moraines are the product of in situ areal stagnation. Clapperton and Sugden (1977) supported this mode of formation but disputed the nature of the environment of deposition and the significance
of stagnation. The apparently irregular, uncontrolled nature of hummocky moraines led to the belief that the stadial glaciers and ice caps stagnated in situ and dissipated by areal stagnation so that all areas were deglaciated simultaneously (Gray and Lowe, 1977; Sissons, 1974b; 1979a). This scenario was thought to be supported by evidence provided by coleoptera (Coope, 1977) which indicates rapid climatic amelioration at the end of the stadial.

This concept of areal deglaciation and simultaneous stagnation conflicts with the view proposed by Charlesworth (1955) and subsequently supported by Thompson (1972) and Eyles (1982; 1983). These authors advocated deposition of hummocky drift during active glacial retreat, with a resultant transverse linear orientation of mounds indicative of successive marginal positions of retreating glaciers. Both modes of formation appear to be reflected in various deposits of hummocky drift. The most common constituent materials comprise boulders of varying sizes in a fine matrix with a wide range of structures that reflect various depositional conditions, such as collapse of the sediment pile or localised flow of saturated deposits. Some stratification may be present and sorting of sands and gravels may also occur, indicative of reworking by running water. Evidence for formation of some hummocky drift by subglacial deformation has also been noted (Hodgson, 1982), and takes the form of longitudinally-oriented 'fluted' ridges in hummocky moraine assemblages.

The primary aim in mapping hummocky drift in the field area was to delimit former glaciers. It was also necessary, however, to investigate the relationship between the distribution of hummocky drift, topography, and other forms of evidence for delimiting the Loch Lomond Readvance glaciers. In addition, the orientation of the mounds was noted where possible, with a view to assessing the mode of formation and thus the nature of deglaciation. Internal sedimentological characteristics and structures within hummocky drift were also noted at exposed sections in order to throw further light on possible modes of hummocky moraine formation, as such structures often indicate differing depositional characteristics (e.g. ice-marginal deposition, subglacial lodgement). Plate 4.2 shows hummocky moraine deposited on a col at NH 9463 in the Monar region.

4.3.1.3 Drift limits.
The distribution of thick drift or drift deposited against drift-free terrain either down-valley or in a valley-side position has also been used, rather more tentatively, to delimit former ice margins. There are three circumstances in which this is important. First, the upper limit of deeply gullied drift has been used in delimiting the former margins of Loch Lomond Readvance glaciers. This criterion has previously been employed to reconstruct former glacier margins on Mull (Gray and Brooks, 1972) and in the east-central
Grampians, the North West Highlands and the Glen Roy area (Sissons, 1974a, 1977c, 1979b). Sissons (1974a, p. 97), for example, states that thick drift plastered on a valley side often has a sharp upper limit that rises consistently upvalley and appears as a small step on the valley side. Substantial deposits of thick till occur in many parts of the field area and in a number of localities the abrupt upslope termination of thick till proved useful for delimiting a former glacier margin, for example the southwards continuation of the glacier limit marked by a lateral moraine at Tomich. Second, former glacial limits have been inferred from contrasts between drift-covered and drift-free terrain, the latter frequently taking the form of upper slopes of bedrock or frost-weathered regolith. This occurs, for instance, south of Achnasheen and on the east-facing slopes of Mullach na Dheiragain in the central part of the field area. Third, a downvalley termination of thick till (especially in conjunction with glacial erratics and boulder spreads) may also represent the former maximum extent of readvance ice. The presence of thick till terminating at a glacier limit indicated by an end moraine has often been cited (e.g. Thorp, 1986). This relationship has been used to infer approximate limits of readvance lobes in some valleys, even when a terminal moraine ridge is absent. Plate 4.3 shows a drift limit to the north of Maiole Lunndaidh.

4.3.1.4 Glacially transported boulders and erratics.
Several previous studies have emphasised the usefulness of boulder spreads for inferring glacial limits (Sissons, 1977c, 1979a; Ballantyne and Wain-Hobson, 1980; Cornish, 1981). Areas of numerous glacially-transported boulders frequently terminate abruptly at limits defined by lateral and end moraines and are often clearly associated with hummocky drift, which may be composed of mounds of such boulders. The frequency of boulders on a valley side is often found to diminish abruptly on the upper parts of slopes above inferred glacial limits, even though the gradient of such slopes remains constant. This evidence was found to be of limited value in isolation, but used in conjunction with that provided by other ice-marginal landforms can provide supplementary evidence. Conversely, where a cover of abundant frost-weathered boulders terminates abruptly downslope, a former glacier limit may sometimes be inferred (plate 4.4).

The number of studies that have employed glacial erratics for delimiting former glaciers is relatively few. Exceptions to this include Sissons (1974a) and Cornish (1981). This type of evidence proved to be of little value in southern Ross-shire, primarily due to the lack of systematic lithological variation amongst the schists that underlie most of the field area.
Plate 4.3.
Loch Lomond Readvance drift limit, north of Maiole Lunndaidh, Glen Ling

Plate 4.4.
Hummocky moraine and boulder limit c. 550m
4.3.1.5 Fluvioglacial features.
On flat valley floors, unkettled outwash spreads are often found within short distances of Loch Lomond Readvance limits (McCann, 1966; Peacock, 1970a; Gray, 1975a; Sissons, 1974a, 1977c, 1979b), whereas kame and kettle terraces and mounds commonly occur just within former glacier margins. Both are used, again with other forms of evidence, to delimit former glacier margins. Eskers were also mapped within Loch Lomond Readvance limits (plate 4.5).

4.3.1.6 Trimlines.
Periglacial trimlines are an additional and important means of delimiting the upper margins of former glaciers. These are considered in detail in section 4.5 below.

4.3.2 Evidence for former direction of ice movement.

4.3.2.1 Ice moulded bedrock.
In the field area, the most widespread form of evidence for determining former ice-flow directions is ice-moulded bedrock. Glacially-modified bedrock is typically smooth and rounded. It has long been recognised as indicative of the existence of former glaciers (Forbes, 1846; McLaren, 1849) and has also been considered a strong indicator of the most recent direction of ice flow (e.g. Bailey et al., 1960; Clapperton and Sugden, 1977; Cornish, 1982). Former directions of ice movement may be detected from bedrock macroforms such as roches moutonnées, extensive glacial pavements, and ice-moulded rock knolls, and by microforms on rock surfaces that retain their glacial polish and which have been modified by ice eroding grooves, channels, striae and friction cracks. Such surfaces are widespread in the field area and are used as an important source of evidence for former ice-flow directions in many valleys (plate 4.6).

4.3.2.2 Striae and friction cracks.
Of the bedrock forms described above, glacial striae have been used most widely in determining the direction of former ice flow. Striae in the Scottish Highlands have been identified and mapped since the middle of the last century (Maclaren, 1849; Jamieson, 1862) and have been recorded since the beginning of the present century by the Geological Survey (e.g. Hinxman et al., 1910; Peach et al., 1913) and subsequently by numerous other researchers.
Plate 4.5.
Kame and kettle topography, Upper Glen Ling

Plate 4.6
Scoured bedrock at 550m above Glen Ling
Recent work (Thorp, 1981a, 1986; Gray and Lowe, 1982; Gray, 1982b) has also stressed the importance of friction cracks for estimating former ice-flow directions. Thorp (1984) listed the different types of friction cracks, which he classified as crescentic gouges and fractures, conchoidal fractures, chattermarks, jagged grooves, and reversed crescentic gouges and fractures. Striae are useful in that they provide a fairly accurate, numerical expression in degrees of the direction of former ice flow. The orientation of friction cracks, can be similarly, if less accurately, determined. Although these features are valuable indicators of ice-flow direction there are problems regarding their use. Care is required in the mapping of striae and friction cracks as markings similar to those formed by the passage of glacier ice can be made by avalanches, drifting shore ice, mudflows and landslides. Differential weathering of some rock types may also result in striae-like forms. A problem that is particularly pertinent to the present study is one related to lithology: on strongly foliated rocks such as mica-schists, schistose quartzites and some psammites, the recognition of striae is most difficult, as the foliated structures of these rocks often give rise to 'pseudo-striae' (Thorp, 1984). The number of unequivocal striae identified in the field was therefore rather less than that which might be expected on more resistant and more massive rock types.

Within the field area, all striae initially mapped by B.N. Peach and his co-workers in the Geological Survey were checked, and some additional sites where striae are present were identified. The identification of friction cracks was, however, largely inhibited by strong foliation of many schistose rocks. Sets of striae were found, in some cases, to cross each other or be oriented differently from those nearby. Most of the striae recorded are believed to be of Loch Lomond Readvance age but in some places it was possible to infer a Late Devensian ice-sheet age for striae on high plateaux. In some cases a glacial limit was clearly evident between two sets of conflicting striae.

4.3.2.3 Roches moutonnées.
Roches moutonnées occur at a variety of scales. Landforms mapped in southern Ross-shire suggest that these features form a size continuum from a microscale of <1m to the macroscale of an entire hill. Problems of interpretation arise, however, due to the fact that the orientation of roches moutonnées may be strongly controlled by structure. Bedding planes, joints, and in particular lineations governed by schistosity and cleavage planes affect the form of both large and small-scale erosional bedforms.

4.3.2.4 Fluted and streamlined drift.
There is evidence in the field area of both a large-scale streamlined sculpturing of the landscape and also fluted and streamlined drift at a much more local scale. Fluted or
streamlined drift has been used in numerous studies in Scotland (e.g. Hodgson, 1982; Ballantyne, 1989) to determine the direction of former ice flow but was important in the present study especially for assessing the direction of flow from corries into the wider valleys.

4.3.2.5 Meltwater channels.
These features have been employed in certain areas of Scotland (e.g. Clapperton and Sugden, 1977) to reconstruct former regional ice flow trends, since their gross alignment often reflects pressure conditions existing at the time of formation, directing meltwater flow towards the glacier margin. However, hydraulic pressure gradients may change during the thinning of an ice mass so that channel directions may cease to reflect that of regional flow lines. In areas of steep relief, therefore, the alignment of meltwater channels must be used with some caution as an indication of former ice-flow directions. An example of a meltwater channel is given in plate 4.7.

4.3.2.6 Erratics.
In addition to providing evidence for delimiting former glaciers, erratics have locally proved instrumental in indicating former ice flow directions (e.g. Harker, 1901; Bailey et al., 1960; Peacock, 1970a,b; Sugden, 1970; Shakesby, 1976; Cornish, 1981, 1982). Although the rock type underlying most of the field area is undifferentiated Moine schist, erratic trains of Torridon sandstone and Cluanie granodiorite have been identified in Strathcarron and Glen Moriston respectively.

4.3.3 Drift types.
Drift masks much of the lower slopes of the field area, but is also present on some hilltops and upper slopes. For the purposes of defining glacier limits and to obtain a detailed picture of the patterns of glacial assemblages it was considered necessary to map different types and thicknesses of drift. Drift cover was therefore mapped under the following categories: (1) regular till sheet; (2) thick (>3 m) undulating or dissected drift; (3) thin (c. 1 - 3 m) undulating or dissected drift; and (4) soliflucted drift. The last-mentioned comprises a smooth mantle of glacially-derived diamicton on upper slopes. In places, the lower limit of soliflucted drift coincides with the upper limit of undulating or hummocky drift, and the contrast between the two marks the upper limit of glacier ice during the Loch Lomond Readvance.
Plate 4.7
A meltwater channel above an outwash fan in Glen Ling

Figure 4.8
Loch Lomond Readvance depositional trimline: lower limit to solifluction
4.4. Periglacial evidence.
A great variety of fossil and active periglacial features are present on the upper slopes of Scottish mountains. The range of such landforms and deposits was outlined in Chapter 2 and it was demonstrated that it is possible to assign most of these to one of two mutually-exclusive categories, namely Holocene (active) or Late Devensian (relict) (Ballantyne, 1984; Table 2.4). Previous research has indicated that almost all relict features occur only outside the limits of the Loch Lomond Readvance (Thorp, 1981a, 1986). It has therefore been considered reasonable to use the distribution of such relict periglacial features as a basis upon which to infer Loch Lomond Readvance limits. There are, however, potential problems with the interpretation of certain periglacial evidence. The importance of the various types of evidence, their relationship to former glacier limits and significance for reconstructing past climates is discussed below. The strategy adopted here was to map all periglacial landforms with a view to distinguishing the distribution of relict and Holocene features, and with particular reference to establishing the location and altitude of periglacial trimlines cut by advancing glaciers during the Loch Lomond Stadial.

4.4.1 Frost-weathered bedrock and mountain-top detritus.
Evidence for the processes of both macrogelivation and microgelivation is present on Scottish mountain tops. Frost-weathered bedrock testifies to the past or present importance of such processes and the resultant regolith is indicative of the dominant type of frost weathering that has operated. Macrogelivation, which involves frost-wedging and frost-shattering, produces coarse clastic detritus. Microgelivation occurs as a form of granular disintegration with a resultant fine debris mantle. Evidence for both is found in the field area. Many bedrock exposures on high ground in southern Ross-shire reveal evidence of being affected by macrogelivation. Such exposures exhibit angular edges, deep open joints and loose angular clasts, partly prised away from the rock in situ by frost weathering along planes of weakness such as joints and bedding planes. The conditions that result in such frost-riving appear to be largely inoperative at the present day, as most rock exposures are covered by mosses, lichens or vegetation. This observation is in agreement with previous studies in the Scottish Highlands, which have demonstrated that the marked contrast in the degree of frost-shattering of rock inside and outside the limits of readvance glaciers (e.g. Sissons, 1967a, 1977b; Thorp, 1981; Ballantyne, 1982) "leaves little doubt that much of the clastic component of Scottish mountain top detritus was produced by macrogelivation under the severe conditions of the Late Devensian" (Ballantyne, 1984). There is some freshly frost-riven debris on Scottish mountains but this only occurs on a small scale, generally at the base of steep outcrops.

Inside the Loch Lomond Readvance limits in southern Ross-shire, bedrock usually appears ice-scoured and evidence of macrogelivation is very uncommon, although some
microgelivation has clearly occurred. Outside the readvance limits evidence of both macrogelivation and microgelivation is common, at least on upper slopes and plateaux. However, the evidence afforded by the appearance of rock outcrops is not always unambiguous. The principal reason for this is that rock structure and texture strongly influence resistance to frost riving, so that in areas of more massive lithology bedrock has been scarcely affected by frost action, even on high ground. The nature of the bedrock and weathered regolith thus reflects not only the processes which have operated, but also bedrock structure. Rock types least susceptible to macrogelivation are those with widely-spaced joints, for example massive siliceous schists. Conversely rock types with closely spaced joints or foliation planes, such as mica-schists, have shattered readily to produce great quantities of frost-riven debris and numerous solifluction forms.

Previous studies have only rarely differentiated between different types of mountain-top detritus. In this study, however, several categories were mapped in order to detect changes in degree of frost weathering. These are shown in Figure 4.4, and are based first on the thickness of the detritus and second on the type of regolith. Ballantyne (1984) identified three types of mountain top regolith (Figure 2.10) and these are described below.

Blockfields and blockslopes are the product largely of macrogelivation of well-jointed rocks. Such deposits are believed to be autochthonous and are most typically developed on the Cambrian Quartzite of the North West Highlands and the Dalradian quartzites of the Grampians (Ballantyne, 1984), but are also present on other lithologies. In the field area they occur, albeit rarely, on more massive Moine schists. Blockfields greater than 1.6 m in depth are known to have developed in the Northern Highlands and fining-down structures have also been observed, a characteristic that has been attributed to heaving (Ballantyne, 1984). In the field, the depth of such deposits was assessed where possible on the basis of the presence or absence of protruding outcrops of intact rock.

Much more common in the field area, however, are frost-weathered diamictons (types 2 and 3 in Ballantyne's classification). In places, periglacial weathering has produced a type 2 regolith with clasts embedded in a matrix of medium and coarse sand and grit. Such regolith is often associated with landforms produced by wind, such as deflation surfaces and wind-patterned ground. Elsewhere plateaux underlain by such regolith are covered by rounded flat-lying slabs with a shallow sandy subsurface layer. Where this type of regolith occurs on slopes, irregular profiles result with rock steps indicating where rock strata have proved relatively resistant to erosion. However, type 3 regolith, with a frost-susceptible matrix consisting primarily of fine sand and an appreciable amount of silt is much more widespread on the hills of southern Ross-shire. This type of regolith is
typical of areas underlain by Moine and Dalradian micaschists and is chiefly characterised by extensive vegetation cover with very regular debris-mantled slopes typically supporting both recent and relict solifluction features.

Three characteristics of mountain top detritus were noted during mapping. First, the depth of the regolith was mapped in terms of one of three categories: (1) thick (0.5m cm) debris mantle with few rock outcrops; (2) incipient (<0.5m detritus cover; or (3) superficial (poorly developed or patchy detritus cover with much bedrock at the surface; (Figure 4.4). Second, in situ plateau deposits were differentiated from those that occupy slopes and are liable to have experienced mass-movement. Third, the type of regolith was assigned to one of the three categories described above.

Many upper slopes, especially those facing west and southwest, are characterised by smooth slopes that are littered with angular frost-riven fragments embedded within a gritty matrix. Gradients rarely exceed 30°. It is not established as to whether the regolith is currently undergoing mass-movement.

Various distinct landforms have developed in mountain-top detritus on the upper slopes of mountains in the study area. These are described below.

4.4.2 Patterned ground
Both relict (Late Devensian) and active patterned ground exists on Scottish mountains. There is a paucity of large, relict features in southern Ross-shire, however, possibly due to the lithology and hence the type of mountain-top regolith. According to Ballantyne (1984), large relict sorted circles and stripes are typically found in type I regolith (block deposits), although in part the lack of such features in the field area may reflect widespread vegetation cover on high ground. Smaller Holocene features are more common on the Moine schists of the field area. Such features as small earth hummocks and shallow patterned ground and were mapped, typically on flat or gentle sloping plateaux such as those of the south Mullardoch hills.

4.4.3 Solifluction features.
A number of solifluction landforms were distinguished during mapping. These were assigned to classes as proposed by Ballantyne (1984). Relict Late Devensian boulder sheets and lobes were identified in some places where type 1 regolith is present. Of these, many exhibit risers up to 2 m high (for example in the Strathfarrar Hills). Vertically-sorted sheets and lobes were also identified on less bouldery regolith, but large, nonsorted, relict solifluction sheets and lobes are most widespread on upper slopes throughout the field area, particularly on type 3 (frost-susceptible) regolith developed on
mica schists. The risers of these sheets and lobes are up to 1.2 m in height. Such features descend to the approximate upper limit of Loch Lomond Readvance glaciers, but do not occur within the area formerly occupied by these glaciers.

Certain active solifluction forms were also identified in southern Ross-shire. Small, active nonsorted solifluction sheets and lobes and especially turf-banked terraces are widespread on the type 3 regolith developed on Moine schist. Such active sheets and lobes were identified by their characteristic low, steep, bulging risers and often complete vegetation cover, and the turf-banked terraces are evident as stripe-like forms with bare treads that trend obliquely across mountain slopes.

4.4.4 Talus and related features.
Talus is the product of rockfall from steep rockwalls. The retreat of the Late Devensian ice sheet exposed many such rockwalls and marks the commencement of talus accumulation in areas that subsequently escaped glaciation. Many rockwalls, however, were reoccupied by glacier ice during the Loch Lomond Stadial and have been subject to subaerial weathering and erosion only during the Holocene, after the wastage of the last glaciers. In consequence, Holocene talus slopes inside Loch Lomond Readvance limits tend to be much less mature in their degree of development than the relict talus that accumulated outside these limits in the interval between ice-sheet deglaciation and the end of the Loch Lomond Stadial (Ballantyne and Eckford, 1984). Talus slopes were therefore mapped as 'relict' or 'recent', on the basis of their maturity, and this difference was used with other evidence for delimiting the extent of Loch Lomond Readvance glaciers.

Protalus ramparts and rock glaciers are known to occur only outwith the limits of Loch Lomond Readvance glaciers, having formed during the Loch Lomond Stadial. No features of either type were found within the field area.

4.5 Periglacial trimlines.
4.5.1 Definition.
A periglacial trimline may be defined as the maximum level to which glacier ice has eroded or 'trimmed' a pre-existing zone of frost-weathered rock or debris on a hillslope. During its advance a glacier will 'trim' underlying bedrock by abrading and smoothing rock surfaces and by removing loose, angular debris up to the level of its surface. Above this level the bedrock remains unaltered by glacial erosion but is subject to frost weathering. It is also possible that drift cover may be present up to the level of the former glacier surface. Thus a trimline created by the most recent advance of glacier ice is represented by the boundary between a downslope zone of glacial drift and glacially-abraded bedrock, and an upslope zone characterised by frost-weathered bedrock that may
or may not support a cover of \textit{in situ} frost-shattered detritus. Where successive glaciers of progressively diminishing thickness have occupied mountain slopes it is possible that two or more trimlines may be identified. This introduces the concept of the existence of two or more weathering zones: as successive trimlines delimit altitudinal zones that have been exposed to weathering processes for different lengths of time, the degree of rock weathering and soil development should be more advanced in the zone above any trimline than that below. In this context the 'trimline' is termed the weathering limit. The idea of trimlines is also closely associated with the 'nunatak hypothesis', which suggests that certain mountain summits remained unglaciated during the last glacial maximum, protruding above the maximum level of the former ice mass (e.g. Reed, 1988).

4.5.2 Trimline evidence.
At a general level, trimlines may be identified on the basis of the distribution of a range of glacial and periglacial features below and above the trimline respectively. Although the term 'trimline' implies a sharp boundary between clearly contrasting surfaces, previous studies have suggested that this boundary is usually present as a \textit{transition zone} rather than a line (Thorp, 1984, 1986) and this was confirmed by evidence in the field area. A consideration of the theoretical glacio-geomorphic formative processes indicates that trimlines or 'trimzones' will inevitably vary in width and in clarity depending on a number of glaciological, geological and topographical variables. Any investigation of trimlines must take these factors into account. Such factors are considered in more detail below.

4.5.3 Glaciological variables.
The amount of debris within a glacier places a fundamental control on whether a trimline will be formed on a hillslope. Clean ice has little erosive power on bedrock surfaces, but given the amount of debris that has been deposited by Loch Lomond Readvance glaciers in the field area, this is not a primary concern here. More important is the degree of concentration of the debris within the marginal zones of the glacier. Abrasion will be greatest where the particle content is high but where ice velocity is not reduced by the viscosity of debris-rich zones (c.f. Sugden and John, 1976). Where there are low quantities of debris in the basal ice a wide transition zone between ice-moulded and frost-shattered bedrock occurs and conversely, where the debris content is more substantial a narrow transition zone may be present (Thorp, 1986).

The width of the transition zone between frost-shattered and ice-moulded bedrock also varies according to spatial and temporal variations in the depth of snow and névé in the accumulation zone above the firm line. Trimlines in this area are likely to be diffuse as the flow vector is directed away from corrie headwalls towards a concave glacier surface. Similarly, in the ablation zone, the margin of the glacier is likely to occupy varying
positions on the valley side and thus the transition zones may also vary in clarity and width (Thorp, 1986).

The thickness of the ice is also believed to be a significant factor in the nature of trimline development. Along the glacier margin, abrasion will be at a minimum near the surface and will tend to increase with depth. This variation in the effectiveness of abrasion is likely to be reflected by bedrock surfaces in the transition zone becoming less abraded upslope.

Rates of abrasion of bedrock are also a function of glacier velocity because the effect of higher sliding velocities is to increase the number of rock particles dragged over bedrock in a given time. Rates therefore decrease or increase depending on whether divergent or convergent flow of ice is locally taking place. Transition zones tend to be widest where rates of abrasion have been reduced by divergent flow and narrowest where convergent flows of ice have occurred. The degree to which glacial abrasion has occurred will also reflect the length of time that the glacier has remained in equilibrium at its maximum thickness.

Theoretically then, the most distinct trimlines with the narrowest transition zones will occur close to the equilibrium line at the margin of glaciers that were very active, were heavily freighted with subglacial debris, and remained at their maximum thickness for a long period of time. Trimlines are likely to be absent or diffuse at the former glacier snout and also in higher parts of the source area.

4.5.4 Lithology and structure

Geological controls on the formation of trimlines have been shown to be of great significance. The clarity of a trimline depends on the response of the bedrock to periglacial and glacial processes. In terms of lithology, the structural and textural characteristics of the bedrock have determined both the extent to which the rock has weathered and are an important consideration in the qualitative assessment of the degree of weathering. Thorp (1984) outlined several criteria upon which to make such an assessment: first, the degree of angularity or rounding of edges of bedrock outcrops; second, the number and depth of open joints and weathering grooves; and third, the amount of granular disintegration.

4.5.5 Trimlines related to the Loch Lomond Readvance.

Whereas a considerable amount of research has been undertaken on trimlines in Scandinavia (e.g. Nesje et al., 1987; Nesje and Serjup, 1988) and Canada (e.g. Ives, 1957, 1958, 1975, 1978), recognition of trimlines formed by glaciers of the Loch Lomond...
Readvance in Scotland has been somewhat belated. The explicit development of this concept was initiated by Thorp (1981), who established a method of systematic mapping of the relevant evidence and used this technique to reconstruct the dimensions of Loch Lomond Readvance glaciers in parts of the Western Grampian Highlands. Thorp (1986) later extended this work using similar techniques in the reconstruction of a major Loch Lomond Readvance icefield across much of the Western Grampian Highlands. Trimline mapping has subsequently been carried out in Skye (Ballantyne, 1989a) and this method has been adopted and adapted for the present study as a principal means of determining the vertical dimensions of the Loch Lomond Readvance in southern Ross-shire (see section 4.5.7). In addition to numerous erosional trimlines of the type mainly employed in previous studies and described in detail above, many of the trimlines in the field area reflect glacial deposition. Such trimlines are marked by the upslope boundary between thick gullied drift and a smooth cover of solifluction debris. The transition is often distinct; gullies originate at the junction of the impermeable soliflucted drift and more permeable drift, and the boundary is often highlighted by a change in vegetation cover. Plate 4.8 (p. 80) illustrates a depositional trimline identified in the field area.

4.5.6 Ice-sheet trimlines.

Recognition of the fact that the dimensions of the Late Devensian ice sheet had been greatly exaggerated (e.g. Sutherland, 1984; Bowen et al., 1986) has been associated with a renewed search for older trimlines on the summits of Scottish mountains. Ballantyne (1984) noted that on mountains in the North West Highlands the transition from glacially-moulded bedrock on lower slopes to a thick debris cover of in situ frost-weathered detritus is not gradual (as might be expected if the altitudinal limits of such detritus was controlled by climate alone), but occurs within a few tens of metres. This suggests that the lower boundary of such detritus represents the upper limit of a former ice surface. Ballantyne et al., (1987) subsequently identified a trimline that clearly predated the Wester Ross Readvance in Torridon, and subsequent research by Reed (1988) confirmed that a high-level trimline occurs on mountains throughout northern Ross-shire. Reed suggested that this trimline separates two weathering zones that occur above a third weathering zone in places where a Loch Lomond Readvance trimline is present.

There are two points of contention regarding this high level weathering limit. First, it is possible that it represents the former boundary between erosive, warm-based ice occupying valleys and lower ground and cold-based ice on mountain summits (cf. Sugden, 1968, 1978; Watts, 1983; Ramussen, 1984). It is possible, however, to counter this argument with the fact that the limit is very abrupt in places (although it is obscured
by downslope mass-movement of regolith elsewhere) In addition, trimline gradients have been found to be consistent with former directions of ice movement (Ballantyne, 1990) and may be continued downvalley by ice-marginal depositional forms such as lateral moraines (Grant, 1977; Nesje et al., 1987, 1988).

Second, the age of the mountain-top deposits in northern Ross-shire is debatable. It has been suggested that the initial development of the detritus on some mountain summits pre-dated the Late Devensian maximum as the amount of macrogelivation on these mountains during the Loch Lomond Stadial appears to have been limited, which in turn implies that a prolonged period of exposure to severe periglacial conditions was necessary for the formation of the thick regolith cover present elsewhere. The presence of erratics in the mountain-top detritus implies that summits above the weathering limit have been glaciated, although not necessarily during the Late Devensian (Ballantyne, et al., 1987).

Despite these controversies, the mapping of high-level trimlines has possibilities for the accurate reconstruction of the dimensions of the last Scottish ice sheet (Ballantyne, 1990). One of the aims of this thesis is to examine the feasibility of this concept by mapping evidence for high-level trimlines and ice-sheet movement on the mountains of southern Ross-shire.

4.5.7 Field mapping of trimlines.
The methodology for mapping Loch Lomond Readvance trimlines detailed by Thorp (1981, 1986) was incorporated into the mapping procedures of the present study. In accordance with Thorp's guidelines, efforts were made to identify as many trimline sites as possible. Spurs were therefore selected for detailed attention as erosional trimlines are most likely to have formed or been preserved at such locations. Mapping of trimlines across successive truncated spurs provides a useful means of cross-checking evidence, as the altitude of Loch Lomond Readvance trimlines may be expected to decline downvalley. Both low- and high-level cols and mountain summits were also mapped carefully to establish the extent of periglacial and glacial modification of the bedrock, and thus to establish the maximum altitude of former glacier cover and to act as a check on trimline altitudes. High-level surfaces were examined for microrelief features that would provide some information as to the degree of frost weathering. Thorp (1984) stressed the importance of distinguishing between plucked and abraded bedrock surfaces on a valley side and this was likewise taken into account during the mapping of trimlines on spurs. The contrast in the depth of open joints inside and outside inferred trimlines (Ballantyne, 1982; Thorp, 1984) was also noted where appropriate.
Trimlines were marked on the field maps as a line rather than a zone. The altitude of erosional trimlines was calculated in the field as the mean of the highest limit of extensive outcrops of ice-scoured bedrock and the lowest limit of extensive frost-shattered bedrock. In some cases it was not necessary to make such a calculation because of the sharpness of the boundary, but often the transition zone occurred over a few tens of metres and the mean value was used. Evidence for Loch Lomond Readvance trimlines was often obscured, however, on grass-covered slopes or steep vertical rock faces, and in such areas it was not possible to identify a trimline of any sort. In addition, trimlines tended to be poorly developed below c. 500 m.

The higher summits were also mapped carefully for evidence of a Late Devensian ice-sheet trimline. Mountain top detritus was mapped as outlined in section 4.4. with a view to identifying different weathering zones. An important criterion for identifying high-level older trimlines is the thickness or maturity of the cover of in situ frost-weathered detritus on plateaux and cols.

One of the major problems in the identification of trimlines in the field area was the frost-resistant lithology. The schists often exhibit only limited evidence of Loch Lomond Stadial frost action, and in consequence many erosional forms, such as roches moutonnées and even striae formed by the passage of the last ice sheet over high ground well above the Loch Lomond Readvance limit, are preserved.

4.6. Summary
The types of landform assemblage produced during ice advance and retreat will be determined by a number of interconnected factors including the manner and rate of advance or retreat, the debris content of the ice and its position of entrainment and topography. The overall distribution of a variety of glacial and periglacial landforms will define the extent of the formerly glaciated area with both lateral (ice-marginal) and vertical (valley side) limits. Inevitably, due to spatial variations within limits as a result of the differing geology, topography and glaciological factors many areas in the field area do not contain clear morphological evidence. It was therefore necessary to map as wide a range of glacial and periglacial evidence as possible to minimise errors in the reconstruction of both glacial limits and former ice-flow directions.
CHAPTER FIVE

The Loch Lomond Readvance in southern Ross-shire:
evidence for the latest glacial readvance.

5.1. Introduction

The former existence of glaciers in the field area is indicated by the presence of depositional and erosional forms left behind when the glaciers retreated. These forms, which include such features as striae, roche moutonnées, end and lateral moraines, trimlines and fluvioglacial features are developed under differing glaciological conditions and thus have differing significance for the reconstruction of former ice margins. End and lateral moraines clearly define the limits of an ice margin, although they may not necessarily define the maximum limit of a particular ice movement, and can represent stages in the decline of a glacier. The temporal implications of many of the features used in glacier reconstructions, for example meltwater channels, are indeterminate but their significance may be indicated by their location, for example, with respect to a proposed ice margin. Despite the limitations of individual pieces of evidence, the picture that emerges when all relevant information is combined can provide a coherent interpretation of the presence of a former glacier.

Whilst it has proved fairly straightforward to determine the lateral extent of former glaciers in an area, the determination of vertical extent is often more difficult through paucity of evidence. The altitudinal extent of the icefield in southern Ross-shire has been determined from the identification of trimlines that have been inferred from contrasts in bedrock, detrital and depositional features within and outwith proposed readvance limits respectively. The vertical extent of the former ice mass has been based primarily on field evidence on 297 spurs, 55 cols and numerous mountain summits and ridges, together with complementary evidence in the form of lateral and hummocky moraines, changes in the thickness of drift upslope and concentrations and spreads of boulders.

The identification of periglacial trimlines is crucial to reconstructing the form of the icefield by supplying the necessary information with which to correlate the marginal evidence for the lateral extent of the icefield in the north, west and east of the field area. The collective evidence for the lateral and vertical extent of the icefield enables the reconstruction of the altitudinal form of the icefield and to infer former ice sheds and flow patterns (see below).

Formerly, it has been possible only to establish limits and glacial characteristics of the ablation areas of icefields (and in some cases corrie glaciers) in Scotland, with a few
recent exceptions (e.g. Ballantyne, 1991; Thorp, 1984, 1986). Attempts have been made, however, to extrapolate limits from ice marginal areas into the former accumulation zones (Bennett, 1991; Bennett and Boulton, 1993b). The use of trimlines in conjunction with other types of ice marginal evidence allows valuable information on source areas and equilibrium line altitudes to be gained. This application is particularly useful in areas such as southern Ross-shire as it is mountainous, highly dissected and contained numerous nunataks.

5.1.1. Nature of the local glaciation identified in southern Ross-shire

It is clear that in southern Ross-shire there is evidence for two separate glaciations: an ice sheet glaciation and a subsequent and less extensive local glaciation. This chapter deals with the most recent glaciation that is believed to be part of the Loch Lomond Readvance age, first because the geomorphological characteristics are similar to those described for Loch Lomond Readvance glaciers elsewhere in highland Scotland (e.g. Sissons, 1977) and secondly because the district is contiguous with ground to the south for which an intense glaciation of Loch Lomond Readvance age has already been inferred (Peacock, 1970; Peacock, 1975; Sissons, 1977; Robinson, 1977). It will be hereafter referred to as such.

5.1.2. Aims

The aims of this chapter are first to outline the evidence for the former transection glacier complex in terms of defining its lateral extent and limits and second the evidence for its vertical thickness. This is followed by an analysis of the evidence for the direction of ice movement and then the patterns of deglaciation. The chapter also considers the spatial distribution of the mapped evidence and its glaciological implications. Having delimited the dimensions of the former icefield and associated glaciers, it has been possible to reconstruct the surface form of the ice mass and calculate contours and ice flow lines in order to establish the glaciological characteristics of the icefield and outlet glaciers.

5.1.3

The evidence is depicted in Figures 5.1 - 5.19 (in folder). Numbers on the maps refer to features discussed in the text. Most of the evidence is presented in map form and the following text focuses on describing the most impressive or crucial evidence, and discussing features that are controversial.
5.2. The delimitation of the lateral extent of the transection glacier complex in southern Ross-shire

The results of field mapping indicate the former existence of 5 independent locally nourished ice masses, the largest of which comprised a major icefield and associated outlet glaciers that occupied almost the whole of the field area, the other four comprising small corrie glaciers. The evidence for these former ice masses is described below.

The former lateral limits of the outlet glaciers that occupied the generally East-West trending valleys of southern Ross-shire are defined in several cases by drift limits against rockslopes or aligned or chaotic hummocky moraine, elsewhere by abrupt changes in the orientation of striae and also by lateral and end moraines. Evidence for the lateral extent of the outlet glaciers of the icefield is described in turn.

5.2.1. Carron glacier

The large, impressive terraces at Achnasheen (NH 1657) are an obvious starting point for describing the extent of the glacier that last occupied the lower parts of Glen Carron, terminating at a lake at the junction of this valley with Glen Docherty and Strath Bran, each of which supported individual glaciers (Sissons, 1982; Benn, 1990). The suite of landforms here was initially considered to be associated with a readvance predating the Loch Lomond Readvance (Sissons, 1982) although this has been questioned by various researchers (Ballantyne et al., 1987; Benn, 1989). The present research confirms that these features were formed by a series of glaciers that contemporaneously terminated at Achnasheen but indicates that these relate to the Loch Lomond Readvance glaciation (Figure 5.11). It is necessary to outline the features associated with this assemblage of landforms as it delimits part of the lateral extent of the Glen Carron glacier. To summarise the evidence for the actual glacial limits the upvalley side of the terrace NH 162573 is clearly an ice-contact slope with glaciofluvial sands and gravels revealing signs of slumping and collapse, implying that the terrace was formed at a time when ice occupied the valley. Fluvio-glacial mounds (including a lochan-filled kettle hole) that occur upslope of and close to these terraces indicate that this glacier terminated slightly further downvalley than the ice contact slope (Sissons, 1982), for example at NH 162574 to the SE where drift mounds and ridges extend up to 260 m. The limit is continued to the SSW, on the southern side of the valley, as a belt of boulders that merges into a lateral moraine 2 - 3 m high and then as a distinct drift limit (above which there is exposed bedrock) to NH 158551. Sissons (1982) traced the extent of this former glacier trending slightly upslope into the tributary valley for about 1200 m as a limit to hummocky moraine (NH 171541). On the north-west side of Strath Carron, similar drift mounds...
were traced only to 245 m immediately upslope of the terrace and the limit was defined no further upvalley.

The present study has confirmed the evidence described by Sissons (1982) and Benn (1990) in terms of the lateral limits of the last phase of glaciation and to extend these limits further upvalley and into tributary valleys of Glen Carron (Figure 5.10). The 'inferred' glacial limit of Sissons (1982) located on the northwest side of the valley is replaced by a drift limit and on the eastern side of the tributary valley the limit is marked by the extent of hummocky moraine. Sissons (1982) noted the existence of hummocky moraine in areas of the main valley up to 10 km to the southwest but made no attempt to delimit the ice mass further in this direction.

On the North side of the Carron valley an indistinct drift limit ascends westwards (320 m at NH 137563) to the col at 386 m where the upslope extent of hummocky moraine indicates that ice passed over from north to south at a maximum altitude of c. 410 m. Incipient mountain top detritus and debris slopes define a limit at 420 m NH 099541 below which is dissected drift. Similarly, a drift limit is apparent on both the northern and southern slopes of Carn Beag, descending from 440 m in the region of Coire Mhic Fearchair (NH 1054) to the 410 m col.

Evidence on the southern side of Glen Carron suggests that ice emerged from two of the three corries to the north east of Moruisg (Figure 5.1). Periglacial trimlines delimit the extent of glacier ice on the northwest slopes of Coreachan Raineach, the lateral limit of which occurs to the east of the stream emerging from Loch Coireag nam Mang (NH 120503). The upslope extent of ice is bounded by a series of mature talus slopes which are probably relict features, suggesting that the corrie was unoccupied by Loch Lomond Readvance ice. This limit descends to the east of the Allt Charagain to c. 400 m at NH 137526 and is thereafter delimited by the upper limit to a suite of moraine ridges and hummocks which cover the col to the south-east of Loch Gowan. There is no evidence for ice having emerged from Creag Coire a' Bhuic during the Loch Lomond Readvance. The configuration of hummocks in this vicinity delimits a lobe of ice that terminated in the region of Mointeach Coire a Bhuic, to the east of which hummocks are flat topped, glaciofluvial in form and composition, and are interpreted as outwash terraces. The limit here is additionally constrained by the occurrence of striae outside the proposed limits which consistently reveal a regional movement of ice to the NE.

5.2.2 Meig glacier

The evidence for the maximum extent of ice in the Meig valley is equivocal: there are three possible locations for the downvalley limit to Loch Lomond Readvance ice (Figure
5.16). First, terminal limits of the former glacier that occupied Gleann Fhiodhaig occur in the vicinity of Corrievuic. A distinct lateral moraine is evident, descending from 270 m on the northern slopes of the valley of Coire a' Bhuic to 200 m at NH 202514. This terminates in a double end moraine either side of a rock knoll to the northwest of Corrievuic. The outer ridge is pronounced although only 1 - 2 m high and the inner ridge is more diffuse. The lateral moraine is extended into the tributary valley as a drift limit to 310 m. A large alluvial fan has obscured possible evidence for the extent of the former glacier on the west side of the valley. A drift limit at 380 m upvalley of this, around the eastern spur of Creag coire na Feola, is consistent with a limit at Corrievuic.

Second, there is evidence for movement of ice downvalley of Corrievuic in the form of striae and ice moulded bedrock and in the valley of the Scardroy burn, chaotic and aligned hummocky moraine is present. On the slopes above Loch Beannacharain, striae up to 320 m are directed downvalley whereas those above this altitude are oriented towards the North East. This collective evidence together with minor ridge at NH 267512 indicates a possible ice marginal position and may also be related to nearby drift limits, at NH 245495 for example. The position of this limit downvalley and its lack of consistency with evidence for the altitude of the ice upvalley suggests that it may be related to a recessional stage during ice sheet downwastage, and possibly correlates with an ice marginal position that occurs in Glen Meinich to the NE, that is apparently non-contemporaneous with the clear ice limits at Achnasheen. A similar explanation may be proposed for the Scardroy Burn valley, where a distinct drift limit descends the lower southern slopes of Creag Loighe, occurring at 240 m at NH 205524 for example.

Between these sites, a third proposed limit is possible at NH 245515 where a series of ridges descend the valley side. There is little on the southern, opposite slopes to corroborate this. The first of these proposed limits is considered to relate to the Loch Lomond Readvance. The absence of drift and of ridges or aligned hummocky moraine in the area outside the proposed limit at Corrievuic affirms this as the Loch Lomond Readvance limit. It is this position which coincides with periglacial evidence for the vertical extent of ice further upvalley.

In the area between Corrievuic and Creag nah-Iolaire, hummocky moraine attests to the former presence of glacier ice and the configuration of the hummocks and ridges suggests that ice actively retreated up this valley. Evidence for recessional positions is discussed in 5.9 whilst the thickness of the glacier confined in the steep sided valley is described in 5.4.

5.2.3 Orrin glacier

The evidence for the lateral extent of glacier ice during the last or local glaciation suggests that a lobe of the Orrin Glacier overspilled from the Monar basin (section 5.2.3.) into upper Glen Orrin. Further source areas for the Orrin Glacier were located in the corries
to the north of Sgurr na Fearstaig and Sgurr Fhuar-thuill in the Strathfarrar Hills and Coire nan Eun (NH 208472). Glacier ice overspilled into Gleann Chorainn whilst a larger glacier terminated three kilometres further downvalley in Glen Orrin.

The terminus of the main glacier is marked by a moraine ridge represented in places by a series of mounds, descending from NH 282472 to NH 296476. This marks the downvalley limit to numerous ridges oriented transverse to the valley, together with associated ice marginal forms including kame terraces (NH 270477) and eskers (e.g. NH 288473). To the east of the proposed limit are areas of flat-topped mounds, composed of glaciofluvial material and these are interpreted as outwash terrace deposits (Figure 5.14).

The continuation of the proposed terminal limit upvalley is marked by fragments of lateral moraine. Lateral morainic ridges are present at NH 284463 and NH 293476 and on the northern flanks of Meallan Odhar (NH 275462). A distinct lateral moraine almost 1 km in length descends the southern slopes of this valley between NH 262451 and NH 270456. Its proximal slope is up to 10 m high and, although much of it resembles a kame terrace, sections reveal coarse unsorted debris, angular to subangular with many large boulders. The step-like appearance may be due to the infill of peat and/or the collapse of the lateral ridge. Such lateral ridges are otherwise discontinuous and are interspersed with drift and rock benches (NH 274463 and 271459) which occur upvalley into the north-facing corries of the Strathfarrar hills. Between fragments of lateral moraine the limit is in places represented by a less distinct drift limit such as at NH 245473. On the north side of the valley the extent of the former glacier is represented by a limit to thick dissected drift or a series of drift benches against a bedrock slope for example at NH 272462 and NH 262468.

The former lateral dimensions of corrie glaciers feeding the main Orrin Glacier are delimited in places: for example as a ridge at NH 230454 or as a line of boulders at NH 224453. Their upslope and upvalley continuation is described in section 5.3.

The limit in Gleann Chaorrain is depicted by distinct, descending drift limits, continuous on the north-western side of the valley and discontinuous on the south-eastern slopes. The position of the terminus of the former glacier is recorded by the existence of a thick band of low moraine ridges on both sides of the valley which is wider and lower on the south eastern side but partially obscured by alluvial fans. Glaciofluvial kames occur immediately outside this limit, and the exit to the tributary valley contains a deep meltwater channel.
5.2.4 Strathfarrar glacier

The Strathfarrar glacier formed the major eastern outlet of ice from the Monar Basin. The ice passing into the upper parts of Glen Strathfarrar was supplemented by that from corries to the south of Sgurr na Fearstaig and Creag Ghorm a Bhealaich. The downvalley, lateral limit to ice associated with this local phase of glaciation is located at NH 294385. The former glacier clearly terminated at an ice marginal lake. The terminus of the glacier is represented by a morainic ridge extending across the valley, amongst a number of rock knolls, from NH 294385 to NH 294384 at Cambussorray. The ridge is lower on its northern portion and in this section is c. 7 - 10 m high, occurring as a mound-like feature here. To the west of Deanie power station at NH 294387 deltaic foresets indicate that the glacier snout was calving into a lake. There is evidence for former higher lake levels around Loch Beannacharan, for example at NH 301385 and NH 304392. Fragments of such features are widespread around the western end of Loch Beannacharan and indicate a former level of c. 125 m.

Plate 5.1  Coire Mhuilidh lateral moraine

Lateral moraines depicting the maximum extent of glaciation are present at 330 m at NH 270371 and at 370 m at NH 262370 with distinct and less distinct drift limits between.
On the northern side of the valley, intensively frost shattered bedrock and periglacial detritus with boulder lobes suggests that the corries to the south east of Sgurr a Coire Ghlas and Carn nan Gobhar were unoccupied by glacier ice during the Loch Lomond Readvance. There is convincing evidence however that a lobe of ice, descending from corries in Toll a Mhuic and Toll Sgaile overspilled into the lower parts of Coire Mhuiulidh. A distinct moraine ridge descends steeply downslope towards the Allt Coire Mhuiulidh and crosses this stream at NH 275406, continues downvalley, descending gently and is continued as a line of boulders to NH 277398 at 350 m. A drift limit to the east of Creag a Bhruic descends between 350 and 330 m the junction of the tributary and main valleys about 1 km from the former glacier terminus.

The maximum extent of ice is defined on the southern side of the valley by the uppermost of a series of lateral moraines that occur between NH 220355 and NH 287375. These occur as low drift ridges or as drift benches and are up to 400 m in length. Where ridges are absent, the limit exists as a drift limit, upslope of which is ice moulded or frost shattered debris (e.g. NH 265365).

5.2.5. Cannich glacier

A series of ice marginal positions are evident in Glen Cannich (Figure 5.17). Evidence in the form of descending drift limits, lateral moraine or cross-valley ridges terminate on or near the valley floor at NH 260335, NH 283334 and NH 300339. The outermost of these occurs as a drift limit on either side of the valley: inside the proposed limit is a preponderance of drift whereas outside, ice-moulded bedrock is widespread. This limit is unimpressive, but can be shown to ascend upvalley to coincide with certain lateral moraines, for example at c. 490 m at NH 273321. The evidence is insubstantial and inconsistent with evidence for the extent of ice further upvalley. It is therefore considered to relate to the last ice sheet. Secondly, a well developed end moraine is present well within this limit near the eastern end of Loch Carrie, where hummocky moraine descends sharply downslope, on the upvalley side of a major rockbar that crosses the valley (and dams the loch). This is believed to relate to a recessional stage, as lateral moraines occur upslope of it.

The readvance limit is considered to be located in the vicinity of NH 282334 in Glen Cannich (Figure 5.17). Here a curving lateral moraine is present on the south side of the valley, extending from NH 284331 to NH 27323. On the northern slopes a band of distinct ridges is discernible trending upslope towards Mam Charaidh.

A lateral limit has also been mapped that extends sharply upslope into Coire Dubh, around the southern flanks of An Soutar and on the slopes above Bealach Bhaca at approximately 580 m. A series of lateral moraines (NH 234346 and NH 242357) show
the extent of ice that spilled from the south into the upper Liatrie valley and further lateral moraines define the extent of ice at a number of points upvalley on this northern side of Glen Cannich (NH 239335 and NH 225334; 640 m and 630 m). Lateral moraines are similarly sporadically located on the south side of the valley. The proposed Loch Lomond Readvance limit is thus defined by ascending lateral moraine fragments at 240 m at NH 288331, 550 m in the vicinity of NH 273319, 570 m at NH 254304 and 610 m at NH 235294.

5.2.6. Affric and Abhain Deabhag glacier

The maximum extent of Loch Lomond Readvance ice emerging from Glen Affric is clearly represented in both the Affric valley and the valley of the Abhain Deabhag. It is most obviously present at Knockfin where there is a distinct end moraine across the valley floor (Plate 5.2). A sharply defined ridge, 2 - 5 m high descends between NH 295267 and NH 302267. It is present on the south side of the valley but as a broad band of drift of lower relief. Downvalley of this limit a series of outwash terraces are discernible.

Plate 5.2
End moraine at Knockfin

The limit on the south side of Glen Affric is represented by a limit to hummocky moraine ridges and mounds extending from 480 m. at NH 255197 to 430 m at NH 267209, where it merges into a discontinuous lateral limit, composed primarily of boulders and occasionally present as a ridge 2 - 3 m high. The ridge becomes more continuous at NH
273206 at 360 m descending downvalley, as an end moraine ranging in height between 1 - 8 m high, to 330 m at NH 286324. From this point there is a continuous, sinuous ridge 4 - 5m high that extends to NH 303253 delimiting the southern extent of the Affric Glacier which becomes less continuous but nevertheless traceable to the terminus at Knockfin Bridge. The southern limit to the Affric glacier occurs as a rather convoluted line on the map in the upper parts of the glen. (Figure 5.18).

On the north side of the Abhain Deabhag a lateral moraine ascends the slopes to NH 263263 where it occurs at 360 m. It is almost continuous but represented by a distinct drift limit where a ridge is not present. The limit curves around Carn Fiaclach but is only present as moraine fragments, for example at NH 274272 and NH 278277. The terminus of the former glacier in Glen Affric is at NH 288283 where there is an indistinct drift ridge to the south of the river and a series of morainic mounds ascending the slopes with an upper limit of 310 m at NH 282288. The Loch Lomond Readvance limit is particularly well represented on the northern slopes of this valley. There is clear evidence for the extent of the glacier from 420 m at NH 272287 to 630 m at NH 182266. A distinct lateral moraine occurs as either a step-like form, for example at NH 242273 where it is 10 m wide with a proximal slope 10 - 15 m high, between NH 232273 and NH 219272 becoming a wider step with a proximal slope of 2 - 10 m. Further upvalley it becomes ridge-like with a proximal slope of up to 25 m and distal slope of 0 - 4 m in places and elsewhere occurs as a terrace-like form 10- 40 m wide. Where there is no lateral moraine, it is possible to define a limit to drift, upslope of which there is no bedrock. The Loch Lomond Readvance limit in the upper parts of the valley is defined by trimline evidence and is described below (Section 5.3).

5.2.7 The Moriston/Doe area

The limit of Loch Lomond Readvance glaciers in Glen Doe and Glen Moriston, as described by Sissons, has been checked and altered. A number of minor adjustments to the limit proposed by Sissons can be suggested for this area: it is proposed that the Loch Lomond Readvance limit in Glen Doe is at NH 180145, downvalley of Sissons’s proposed limit at NH 184144. The termini of the glaciers of Coire Sgreumh and Coire Mheadhoin can be extended by 600 m and (on the south side of the valley) 400 m respectively. There is also evidence that the corries of the upper part of Gleann Fada supported ice and that it is possible to define a downvalley limit at NH 168178. Lateral moraines for this glacier identified at NH 106188, NH 132187 and NH 102160 depict its lateral limits.

The maximum extent of the main Moriston Glacier, described by Sissons (1977), was remapped and its location judged to be accurate. The well-defined moraine that occurs
on the southern side of the valley is clear evidence for the Loch Lomond Readvance limit, together with the upslope extent of (in places) numerous moraine ridges and hummocky moraine.

At the southernmost extremities of the field area, the limit mapped by Sissons was checked, as far as it goes, and was considered to be an accurate representation of the readvance limit. Sisson's mapping ends at the northing 10 (Ordnance Survey). The present research continued mapping to the west: the ridges revealing the maximum extent of ice, that are present on the south side of the valley, for example at NH 255106, are continued to the southwest into the valley of Loch Loyne.

5.2.8 Duich glacier

The evidence for the former extent of the lower parts of the glacier that flowed towards Loch Duich is scarce and, due to this (in part), equivocal. The geomorphological mapping indicates three possible downvalley positions for the maximum extent of the western Glen Shiel glacier: first, there is evidence in the region of Nostie for ice-contact deposits at NG 858270, suggesting that this is either the location of the Loch Lomond Readvance limit or a deposit associated with a large valley glacier that occupied this area during the downwastage of the last ice sheet, either relating to a readvance (Wester Ross Readvance age) or an undefined recessional stage (Figure 5.3). There is little corroborative evidence in this vicinity, for a Loch Lomond Readvance maximal limit.

Further glaciofluvial deposits are located at Dornie. These may have formed at the maximum of the Loch Lomond Readvance when the ice termini may have been located at the junction of Loch Duich with Loch Long. Evidence in support of this is lacking here. A number of ill-defined, gently descending drift limits do indicate, however, that a glacier terminus may have indeed occurred in the vicinity of Dornie: drift limits are apparent at 120 m at NG 887258, 200 m at NG 892252 but an outermost limit appears to be represented by a lateral moraine upslope of these between NG 889259 and NG 900250 at 320 m descending to 230 m. There is a lateral moraine, occurring as a thickening of drift in places, a bench like form, ascending from 210 m at NG 894250 to 310 m at NG 902248. This is continued as a drift limit around Boc Beag at 350 m, although not always as a bench. A distinct melt water channel at 420 m is more likely to represent a slightly higher limit. The only visible drift limit on the south side extends from 140 m at NG 880237 to 210 m at NG 884230. There is more convincing evidence for the extent of the readvance limit further upvalley in the form of trimlines which are consistent with the interpretation of the features at Dornie as representing the readvance limit (section 5.3). In addition to the geomorphological evidence for the main valley, the tributary valley of Coire Dhuinnid was mapped. The distribution of hummocky moraine, in association
with pronounced lateral moraines and boulder spreads indicates that ice passed out of the valley at an altitude of 340 - 330 m. A distinct lateral moraine and drift limit exists at 470 m at NG 943252.

6.2.9 Glen Lichd glacier

Ice flowing west in Glen Lichd, fed by numerous corries on the northern side of The Five Sisters and the Glen Shiel Hills, entered Strath Croe and thereafter flowed into the area now occupied by Loch Duich to the South and also spilled over a col between Sgurr an Airgid and Beinn Bhuide. Some of the Glen Lichd ice moved northwards across the eastern flanks of Beinn Breac, to the west of A Glas Bheinn. The western extent of the Glen Lichd valley glacier was at its maximum, beyond the western end of the valley. There is evidence, however, for post maximum recessional limits (cf. section 5.7). The altitudinal extent of this glacier is discussed in section 5.3.

5.2.10 Upper Glen Affric/ Gleann Gaorsiac

Similarly, the area comprising the eastern part of Glen Affric was well within the limits of the maximum extent of the Loch Lomond Readvance, and recessional stages pertaining to the former extent of ice present in this area reveal only a minimum lateral extent for the Loch Lomond Readvance maximum. There were, however, several nunataks as identified largely from trimline evidence (5.3).

5.2.11 Glen Elchaig

On the basis of geomorphological mapping in this valley two possible readvance limits are proposed. The former position of a glacier snout is indeed represented at numerous locations within this valley. Ridges and mounds associated with recessional stages occur for example for 2 km west of Loch na Leitreach at NH 020275, NG 983274 and NG 972277. There is, however, a more impressive end moraine on the north side of the valley at NG 956285 which probably marks the maximum extent of this Loch Lomond Readvance glacier. The end moraine comprises a series of hummocky mounds that extend steeply upslope assuming the form of a moraine ridge. The limit curves to the east for at least 600 m and ascends the valley slope as a drift limit at a rather gentler angle to 150 - 160 m at NG 973285. A meltwater channel occurs outside a distinct lateral moraine (4 m high) at NG 960286 and, outside the proposed limit, smoother debris slopes occur. There is little that attests to a former glacial maximum on the southern side of the valley at this point, except an indistinct drift limit that ascends the steep slopes to
the crag of Creag Loisgte (NG 964277). Elsewhere, the steep rocky slopes have preserved no evidence. Downvalley of this proposed limit slopes, immediately outside it (NG 955295) appear rather frost shattered with open joints and spreads of debris below outcrops.

An ice marginal position is also recorded further downvalley where a well-defined moraine ridge extends from NG 912304 to NG 907302 on the north side of the valley, to the west of Sallachy. There is a corresponding, although less impressive ridge on the south side of the valley which is continued immediately upvalley by a distinct drift limit, above which there is exposed bedrock and in places this defines a limit to hummocky drift. In the light of the mapping in adjacent valleys, the evidence for sea levels and the trimline evidence for the upper part of this valley, this limit is considered to be outwith that of the Loch Lomond Readvance, which it is thought, is that located at NG 956285.

5.2.12 Glen Ling

The position of the ice maximum in Glen Ling is equivocal. Several ice-marginal positions are discernible in the valley: first, a drift ridge descending the slope at NG 944333 appears to be an ice-marginal moraine; second, there is a moraine ridge descending the hillslope at NG 958327 to the southwest. The moraine marks the approximate limit to undulating drift. The extent and maturity of the relict talus slopes at NG 966334 on the south side of the valley however, suggest that this area lay outside the limits of the Loch Lomond Readvance.

A suite of landforms occurs in the vicinity of NG 962339. In general the morphology of this glen suggests three phases of depositional/erosional development: First, Deposition of stagnant ice deposits including kames, kettles and high glaciofluvial outwash terraces, formed during the retreat of the last ice sheet. Secondly, Extensive planation by vigorous meltwater streams subsequently occurred, probably during the Loch Lomond Stadial to produce an extensive unkettled terrace. The truncation of contorted outwash by an intermediate terrace at NG 965338 suggests that this intermediate feature formed mainly by erosion of earlier deposits. Thirdly, Further Holocene incision with low terrace formation and the lowering of the river to its present level (Figure 5.20).

There is strong evidence in the intervening valley between Glen Ling and Glen Elchaig for a small corrie glacier having occurred on the northern slopes of Sguman Coinntich. At NG 971310 a moraine ridge 150 m long extends ENE - WSW. Its proximal height is 5m and the distal edge is 3.5 m high. Exposures in the ridge, e.g. NG 977309 reveal boulders embedded in a sandy matrix. The limit is continued upslope on the downvalley side by a drift limit. Larger glacially transported boulders occur inside the ridge with
Figure 5.20  
Depositional landforms in upper Glen Ling.
melt-out hollows near the ridge crest and at the western end of the ridge. Boulders can be traced from the backwall and thus define the extent of the former glacier (NG 977308). Periglacial debris extends right up to the distal slope but is absent inside the apparent limit. Mature debris mantled slopes NG 965310 suggest that the valley did not support Loch Lomond Readvance ice and this supports the evidence described above in terms of delimiting a Loch Lomond Readvance corrie glacier.

In upper Glen Ling the limit is expressed as a drift limit with a more pronounced bench of drift (NG 989388 at 290 m).

5.2.13 Strathcarron glacier

The maximum extent of the Loch Lomond Readvance in Strathcarron is fairly certain. A lateral moraine descends the slopes from east of Cnoc na h-Atha to the end moraine extending straight across the main valley from NG 946444 to NG 957435. It is a composite feature with two major ridges trending across the valley, the innermost of which is of a lesser height. There are also subsidiary mounds and ridges. The outer ridge has a broad undulating crest with a maximum height of 15 m. In its eastern sections it is lower and grades into kettled topography and an alluvial fan. Inside of these two principal moraines are a series of lower and less pronounced moraine ridges, interspersed with extensive kame and kettle topography including eskers, kettle holes, small and flat-topped mounds of glaciofluvial material. Sections such as at NG 948444 reveal ice contact deposits with horizontally bedded sands and gravels that have been subjected to collapse and disruption of the strata. This suggests the close proximity of the melting ice. The moraine limit continues along the valley side as a broad system of linear mounds, rising north-eastwards to the break of slope.

A distinct drift limit in places evident as a lateral moraine, (notably at NG 962437, NG 970440, NG 974443 and NG 980445) can be traced upvalley on the southern side of Strathcarron, rising, at first, gently then at NG 990453, ascending steeply as a lateral ridge up to 420 m. The limit then curves around the eastern slopes of Carn Mor as a drift limit. Its downvalley extent in Coire Taodail is uncertain but probably marked by a series of morainic mounds and plug of drift at NG 997445.

On the northern side of the valley, the uppermost limit to lateral moraine ridges is at approximately 190 m on the south west slopes of Cnoc nan Each. This is continued up the Glen Carron valley as a discontinuous drift limit, locally represented as a ridge or bench, to NG 982472 where it trends NW at the western extremities of Coire Dubh.
5.3 The vertical extent of Loch Lomond Readvance ice masses in southern Ross-shire.

5.3.1.

The altitudinal extent of the Loch Lomond Readvance icefield in southern Ross-shire has been determined from the identification of trimlines that have been inferred from contrasts in bedrock, detrital and depositional features within and outwith proposed readvance limits respectively (Section 5.1).

Table 5.1 gives the mapped or calculated trimline altitudes these are shown in Figures 5.21 and 5.22.

5.3.2 The vertical extent of the ice across the field area

Within the area of the southern Ross icefield periglacial trimlines define the extent of several nunataks. The former ice surface can be in many places thus precisely defined especially towards the centre of the icefield. This evidence complements limits based on end, lateral and hummocky moraines, the limit of thick drift and the ice contact slope of glaciofluvial landforms. This has enabled lateral and vertical limits to be combined to provide a detailed diagram of the Loch Lomond Readvance limits for southern Ross-shire. The correlation of eastern and western limits is essential to acquiring an understanding of the form and glaciological characteristics of this part of the icefield. This has also been achieved for the area around Rannoch Moor and for Skye (Thorpe, 1984,1986; Ballantyne, 1991) but elsewhere such information has to date been lacking.

The altitude of individual corrie glaciers is determined through a combination of trimline and depositional evidence. To the north of Sguman Coinntich, the corrie glacier described in 5.2 is defined in its upper portions by a periglacial trimline at NG 976306 that descends from an altitude of c. 800 m to 630 m and terminates in a moraine at 410 m. Elsewhere, in the numerous cases where corrie glaciers feed into valley glaciers, the upper limits of corrie glaciers are identified on corrie backwalls and sides. Such maximum altitudes are shown in Figures 5.21 and 5.22. For example in Glen Shiel the upper extent of corrie glaciers is delimited by trimlines such as at NH 011102 where there is a lower limit to frost-shattered slabs on the backwall, below which the bedrock is distinctly ice-moulded in a downvalley direction. A periglacial trimline marks the extent of ice in NH 0219.
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The maximum altitude of valley ice in the field area is at 820 m at the ice shed in Glen Shiel. The maximum former altitude of the ice in the valleys towards the north of this decreases with maximum values of 550 m and occurring in Gleann Fhidhaig. The maximum altitude of the ice field in Glen Carron is more readily defined by lateral moraines as it lies around the altitude where periglacial effects are diminishing the extent that trimlines are indiscernible.

Nunataks occur within the limits of the icefield and include the summits of the Carn Eighe Massif, An Riabhachan and Sgurr na Lapaich, Moruisg, Maoile Lunndaidh and Sgurr nan Ceannachain. Breaches are uncommon in the area as there are few low points in the east-west trending ridges. Trimline altitudes decrease eastwards from the central ice shed identified in 5.22 and tend to merge into lateral moraines to define the discrete outlet glaciers. This pattern is also reflected, although less clear, towards the west.

**5.4. Ice directional evidence for the Loch Lomond Readvance**

Several forms of evidence attest to the direction of former ice movements of the Loch Lomond Readvance in southern Ross-shire. Erosional evidence includes striae, ice-scoured bedrock forms, friction cracks, crescentic gouges and roches moutonnées. Streamlined drift and erratics deposited by the ice sheet and the alignment of glacial troughs and breaches are also evident as ice directional indicators.
Figure 5.22
Trimline altitudes in the south of the study area

Figure 5.23
Loch Lomond Readvance Straie

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[Map of Trimline altitudes in the south of the study area]

[Map of Loch Lomond Readvance Straie]
Striae and roches moutonnées found on the Geological Survey 1: 63360 and 1: 50,000 scale sheets were checked in the field and plotted onto 1: 25,000 Ordnance Survey maps together with those derived from field mapping. Such striae together with friction cracks and strongly streamlined bedrock forms are shown in Figures 5.23 and 5.24. This diagram also shows the direction of former ice movements indicated by roches moutonnées. Former ice directional movements derived from this evidence placed within the limits of the Loch Lomond Readvance are depicted in Figure 5.25.

The evidence for former ice movement within the limits of the Loch Lomond Readvance may be summarised with reference to particular areas. Striae and ice moulded bedrock show that ice flowed from the valley of the Allt a Chonais (NH 070480) south into Glen Carron and further such evidence in Glen Carron suggests that the ice diverged to flow east and west. In addition to the main glacier flow to the east in Glen Carron, striae and ice moulded bedforms reveal that there was a flow of ice from two corries to the NE of Moruisg. The presence of widespread frost shattered bedrock in the third at NH 127505 indicates that this was ice free during the Loch Lomond Stadial. There is little evidence in the corrie on the western slopes of Moruisg to indicate ice movement through it but a lateral moraine is present which suggests it was probably occupied.

Although much of the valley of Gleann Fhiodhaig is covered in drift, the lower part which was occupied by readvance ice contains evidence suggesting that ice moved along the valley. Downvalley directed striae on lower slopes contrast with those with a more easterly orientation immediately upslope. To the south, the main valley in upper Glen Orrin is devoid of ice directional forms, although fluted moraines are oriented out of the corrie of An Gorm Loch at NH 250461, together with striae and ice moulded bedrock further towards the backwall. Striae close to the backwall indicate that it was in its westernmost part almost fully occupied by glacier ice whilst to the east of this the corries were only partly utilised. Ice moulded bedrock and depositional forms suggest that ice may have moved through a small corrie glacier to the E of Creag Ghorm a Bhealaich (NH 250435). Evidence also indicates that ice emerged from the westernmost corries on the south side of these Strathfarrar hills, but not from adjacent corries to the east of Sgurr a Choire Ghlaist and Carn nan Gobhar. The corries of Loch Toll a Mhuic and Allt Toll Sgaile were partly occupied as shown by striae, fluted moraines and ice moulded bedrock, features which are also present below the corries and which indicate a downvalley flow of ice. Apart from this, ice directional landforms show a clear flow of a major ice stream from the Monar basin into Glen Strathfarrar. Such features occur for several kilometres downvalley and beyond the readvance limit, especially on the rock bars that cross the valley, although these are clearly not necessarily of Loch Lomond stadial age.
Strongly ice moulded bedrock oriented downvalley occurs on the northern and southern sides of Loch Mullardoch in Glen Cannich. This trend occurs downvalley of the loch where the slopes are not covered in thick drift. To the south, striae and ice moulded bedrock reveal that the eastward flowing glacier in Glen Affric terminated East of Loch Beinn a Mheadhoin and a broad distributary crossed the intervening ridge into the Abhain Deabhag valley to reach the readvance maximum at Knockfin. Ice originating in upper Glen Affric spilled across a 610 m col (NH 130177) and into the upper reaches of the Allt Garbh valley. Similarly well-defined ice moulded bedrock shows that ice emerged out of the corrie at NH 014106 and that ice moved E-W along Glen Shiel at NH 0210: the west side of outcrops are more craggy than ice moulded the eastern side. Ice moulding in NH 0219 suggests that ice moved out of the corrie to the North then swung round to NW.
A northward flow of ice on the North side of Glen Affric ice across the ridge to the east of Sgurr na Lapaich (153243) into Gleann nam Fiadh is indicated by striae. Ice from Glen Shiel flowed to the W and SE of Carn a Choire Ghairbh into the Gleann na Ciche and the upper reaches of the Allt Garbh respectively and thereafter entered the valley of the upper Affric valley. Ice from the Loch Cluanie basin crossed a 730 m ridge in a north-easterly direction into the R. Doe valley, carrying with it Cluanie granodiorite (Figure 2.10).

Glen Shiel reveals the most impressive evidence for the direction of ice flow. This occurs largely in the form of roches moutonnées, striae and ice moulded bedrock, but some directional depositional forms occur, for example trending ENE just east of the watershed in this glen. The corries on the north side of the Cluanie ridge all reveal former flow of glacier ice into Glen Shiel. Similarly, those on the northern side of Glen Shiel were occupied and exhibit ice directional features to some extent. Excellent evidence is also located in the corries above Glen Lichd, which appear to have been extensively occupied.

Features indicating a westerly movement in Glen Shiel are similarly clear to the west above and upvalley of Loch Duich and in Glen Lichd. The evidence becomes more diffuse on the plateau between Loch Duich and Glen Elchaig, but the direction of former flow becomes clear in Glen Elchaig, contrasting markedly in direction with evidence relating to the Late Devensian ice sheet. Further north in Glen Ling and Strathcarron it is more difficult to separate ice directional forms in terms of the Late Devensian and Loch Lomond Readvance.

Ice directional evidence towards the central source areas of the glaciers show that most of the corries in the central massifs contributed to the large east and west flowing ice masses. Included are those of the Sgurr nan Ceathreamnán-Carn Eighe range, although corries to the east of Carn Eighe and Mam Sodhail were only partly occupied. Ice also flowed from corries surrounding an Riabhachan and Sgurr na Lapaich, although to the east of this, around Carn nan Gobhar, more corries were unoccupied. Similarly, in the vicinity of the Killilan Hills, striae and ice moulded bedrock show that corries were not fully utilised, and some supported only small corrie glaciers.

Almost all of the corries surrounding the Monar basin reveal evidence for former ice flow into the basin. This includes those north facing corries of Aonach Buidhe, Bidean a Coire Sheasgaich in addition to the north facing corries of the An Riabhachan range. Many south-facing corries did not support glacier ice.
A generalised map of ice flow patterns for the Loch Lomond Readvance is given in Figure 5.25. It is probable that the pattern of striae and associated ice-erosional features is the product of ice flow patterns occurring at successive stages during the active decay of the ice (Section 5.9). The pattern may reflect changing flow directions. The evidence for the direction of former ice movement indicates that the dominant pattern of ice flow was to the East and to the West from a central source area.

5.5. Characteristics and glaciological implications of the geomorphological evidence

The morphological, sedimentological, structural and spatial characteristics of glacial and periglacial features of Loch Lomond Stadial origin are discussed below. Features that delimit the readvance limit are described first. Landforms and deposits occurring within these limits are then considered, followed by a discussion of other related geomorphic evidence.
5.5.1 End moraines

Well developed end moraines delimit the former maximum extent of only a few of the transection glaciers. Pronounced moraine ridges occur in the eastward-trending Glen Moriston, Glen Affric and upper Glen Orrin, whereas a lack of distinct moraine ridges at maximal positions for western oriented glaciers is noted, as in Glens Shiel and Ling, although the substantial moraine ridge in Strathcarron is an exception.

The reason for such contrasts between West and East is uncertain. A number of explanations are plausible: first, the relative proximity of western glaciers to the coast may be significant. In the same way as the former glacier termini of the Creran and Linnhe glaciers to the south were tidal, so mapping indicates that the Glen Shiel glacier terminated in Loch Duich and Loch Long. In this instance, and perhaps others, it is likely that such deep water fjords conceal the presence of a moraine ridge (or series of ridges). Recent seismic and scanning surveys have located such submarine ridges in western fjords, including Loch Duich (J. Dix, pers. comm) but the results of such investigations have to date proved equivocal. Alternatively, the lack of end moraines in deep water on fjord floors may suggest that glaciers failed to reach a sufficiently long steady state position for end moraines to form. (Mercer, 1961; Paterson, 1981; Thorp, 1991).

The E-W variation in the incidence of end moraine could feasibly be the result of palaeoclimatic differences across the area. Evidence suggests (Section 5.9) that climatic conditions in the west were more unstable and the ice front was not stationary at its maximum position for a long period of time. Thus, glaciers that were sensitive to minor climatic fluctuations would fail to produce a substantial end moraine as terminal positions would vary considerably. Alternatively the associated variations in topography (the effects of steep sided valleys) and glacier dynamics may account in part for the preferred formation of end moraines in the east.

The pattern of moraine distribution contrasts with that described for Applecross, to the north of the present field area where terminal moraines are more pronounced to the west and south. In the absence of other decisive factors, Robinson (1977) attributed the preferred incidence of moraines to aspect, and therefore to climatic control.

It has been suggested for other areas (Thorp, 1991) that western glaciers, flowing largely below 600 m (and therefore with higher rates of ablation than those in the east), may have generated appreciable quantities of meltwater (Flint, 1971; Gustavson and Boothroyd, 1982), and may have resulted in debris having been entrained by meltwater, thus preventing moraine formation. This assertion may be applicable in southern Ross-shire.
The composition of end moraines and the internal structures that have been examined clearly reflect differing environmental conditions that existed throughout the area during the Loch Lomond Stadial. The fact that the glacier in Glen Strathfarrar terminated in a proglacial lake is revealed by deltaic structures in the fine-grained silts and sands observed at several small sections in end moraines near the eastern end of Loch Beannacharan. However, in most of the terrestrial termini, glacial diamict is mixed with glaciofluvial sands and gravels, for example at Knockfin (where recent building developments exposed deposits), Glen Carron and Glen Orrin, reflecting varying local conditions at the time of deposition. These characteristics are also present in recessional moraines (5.5.4).

5.5.2. Lateral moraines

Lateral moraines occur in the lower parts of almost all the eastern valleys that supported outlet glaciers. This preponderance of lateral moraines in the eastern valleys is not replicated for the west of the area. Here evidence in the form of lateral moraines is of a much more fragmentary nature. They do however occur in upper Glen Ling. It is noted that a similar east-west trend also occurs with respect to lateral moraines associated with recessional stages. Topography has an obvious control on the occurrence and preservation of lateral moraines. The valleys in the west are generally steep sided and this may have restricted moraine formation. Further, lateral moraines are preferentially present on gentler convex slopes which in this area tend to occur on north-facing slopes.

Sedimentologically, lateral moraines were observed to differ markedly, composed of varying proportions of sands and gravels and diamictons, often with collapse structures from slumping as the adjacent glacier partially supporting the sediment downwasted.

5.5.3. Drift limits

Drift limits define many kilometres of the limit of outlet glaciers in southern Ross-shire often occurring between more distinct features that are readily identifiable as moraine ridges, hummocks or benches (Plate 5.3). They occur as a more or less abrupt change in the thickness of drift upslope or below bedrock. This form of glacial limit is particularly well developed in the eastern part of Glen Carron and in the upper northern section of Glen Ling. Drift limits also occur in tributary corrie glaciers, for example in the uppermost 4 km of Gleann nam Fiadh. Although mapped as a drift limit, some forms may be simply unpronounced morainic features, reflecting changes in basal glaciological conditions. Boulder limits occur in several locations.
Abrupt limits to arcuate spreads of boulders contrasting with boulder free areas outside the limits occur in Toll a Mhuic, the corrie west of Sgumman Coinntich. A high level lateral moraine and drift limit is located at NH 169460 (Plate 5.4).

Plate 5.4
High level lateral moraine and drift limit
5.5.4. Hummocky moraine

The area of southern Ross-shire contains diverse forms of hummocky moraine. Many areas of valley sides and floors are covered with hummocky moraine which contrasts strongly with the upper hill slopes that consist largely of bare ice scoured bedrock surfaces. Most hummocky moraine in the field area is of low relief (<5 m) although the form and morphology of the hummocky moraine is clearly variable. The size of individual mounds varied between 0.5 and 5 m in height. Conical or roughly equidimensional mounds tended to be between 2 and 10 m across but often took the form of longer ridges: continuous ridges of up to 2 km were identified and often the configuration of lines of mounds represented ridges that had been dissected. In other places, low mounds, with negligible relief are separated by hollows often infilled with peat. Elsewhere they exist as kettled forms.

Although the distribution of mounds was in many cases apparently random, producing a chaotic drainage system and a disordered undulating landscape, alignments may often be discerned in the mounds. In places the moraine occurs in distinct lines or chains which appear to represent former end - positions of a valley glacier. Here the moraines are characteristically sharp-crested, 2 - 5 m high and often undulating in form. The commonest orientation of mound crests is in an oblique downvalley direction, producing a chevron pattern where moraine occurs on both sides of the valley. In some case the orientation is 90° to the valley axis (cross valley) with impressive suites occurring in the eastern valleys, especially in Gleann Fhiodhaig and Glen Affric. It has been suggested that chaotic hummocky moraine is rare, but there are numerous areas of hummocky moraine which are considered to exhibit no orientation. Such occurrences are primarily in the upper parts of mapped valleys especially at the junction of corrie and valley glaciers or at the exit of corries (Plates 5.5 and 5.6).

Several factors clearly control the distribution of hummocky moraine. Most hummocky moraine occurs in the lower valley and valley floor for example in Glen Carron although some are formed at high levels such as that occurring in corries surrounding An Riabhachan. The abundance of moraines in the valleys towards the north of the field area contrasts with their much sparser development to the south. This may be related to glacier size.

Differences in the size and form of hummocky moraine may relate to topographic features. Smaller, ill-defined ridges occur on valley sides where the gradient is steep compared to the highly pronounced forms on valley floors, a trend similar to that found in other areas (Sissons and Grant, 1972; Thompson, 1972; Thorp, 1984). A higher incidence of moraine ridges on the northern sides of many of the E-W trending valleys is
Plate 5.5
Low morainic hummocks and ridges near Loch a Chilaidheihm at 550m, Monar (NH152465)

Plate 5.6
Hummocky moraine on col at 490m, Monar NH150463
noted especially in deeply incised valleys that probably experienced strong insolation contrasts.

Some instances of hummocky moraine correlate with large inputs of rock debris onto the surfaces of corrie and valley side rock walls by means of snow and rock avalanche and other periglacially induced mass movements as described by Sissons and Grant (1972), Sissons (1967a), Eyles (1983), Benn (1989a). The abrupt termination of boulder lobes and solifluction at glacial limits suggests that a great deal of supraglacial debris was derived from snow free slopes undergoing severe periglacial weathering (Sissons, 1976; Ballantyne, 1984; Thorp, 1984, 1991).

The sedimentological characteristics of ‘hummocky moraine’ was determined for a number of sites. Exposures were analysed in Glen Elchaig (Plate 5.7), Glen Ling and Glen Affric. Many hummocks were boulder studded with clasts up to 1 m. Water sorted silts, sands and gravels occur. In other localities a coarse diamicton was observed with more or less matrix. Boulder spreads between hummocks were observed frequently such as in Glen Affric and Gleann Fhioddaig. The most common sedimentology was a bouldery diamicrt with a sandy matrix. It is difficult to comment on the genesis of such occurrences due to their sporadic nature and to the variability of the sediments and possible origins. In a few instances reworking was noted. However, on the evidence available, it is possible to state that the mode of origin varied with both evidence for ice marginal formation as well as for stagnation deposition.

Plate 5.7
Fluvio-glacial deposits in hummocky moraine belt, Glen Elchaig
It has been argued that where 'transverse' ridges (or assemblages of hummocky moraines) occur, they are certainly ice marginal forms as the only transverse landforms of modern glaciers are ice marginal moraines (Sharp, 1984; Boulton, 1986; Bennett, 1991). It has also been proposed that ice marginal moraines are asymmetric (Matthews et al. 1979; Matthews and Petch, 1982; Shakesby, 1989). The sedimentological and morphological evidence (Figure 5.27) of the transverse moraines in the field area does support the hypothesis that the forms are ice marginal (push moraines, dump moraines and outwash fans), and that active retreat occurred where these are found.

Figure 5.26
Transverse Moraines

5.5.5 Glaciofluvial deposits

Many of the hummocky forms described above are undoubtedly of glaciofluvial origin. However, it is difficult to differentiate between individual mounds or ridges, from their morphology, those that may be composed of glaciofluvial sands and gravels and those that may be composed of glacial till. In fact, many such forms may contain elements of both types of deposited material (Plate 5.7).
Apart from hummocky glaciofluvial landforms, there are distinct suites of other glaciofluvial landforms. Evidence for glaciofluvial deltas occur at NG 938310 and NG 947282 inside the maximal limit and are associated with the declining Elchaig glacier with meltwater dammed in the first case in the Ling valley against the glacier surface, unable to drain and in the second instance, in an enclave between the glacier and valley side.

Ice contact outwash terraces and fans occur in association with the maximal limits at Dornie at the mouth of Glen Shiel and emerge from the proposed limit in other valleys such as Glen Orrin, Strathconon and Glen Ling. Outwash is also clearly present within the maximal limits and often terminate upvalley at identified recessional moraines (Plate 5.8).

**Plate 5.8**

*Outwash fan in Glen Ling from its apex*

A suite of glaciofluvial forms which includes eskers and kames is evident in Glen Orrin. The eskers are generally <300 m long but the kame terrace at NH 270477 is more impressive. Sections reveal poorly developed sands and gravels with some clay lenses. This landform assemblage suggests a stagnating ice mass. The reasons for this are discussed in 6.7. Other occurrences of eskers are at the western end of Glen Elchaig and in the Abhain Deabhag valley north of Glen Doe but glaciofluvial forms are not common and attest to the fact that the mode of deglaciation was not generally by widespread stagnation but by active retreat (Eyles, 1983; Benn, 1991; Bennett and Boulton, 1993).
**5.5.6 Ice dammed lakes**

Glaciolacustrine sediments occur in Glen Strathfarrar and Strathconon where the former glaciers terminated in a lake, evidence for which is discernible above the level of the present lakes. It is also proposed that in addition to the ice dammed lakes that existed in Glen Doe (NH 250120; Sissons, 1977d) and at Achnasheean (Sissons, 1982; Benn 1989a), an ice dammed lake may have occurred in the middle sections of Glen Ling at NG 957334. These features indicate that ice retreated down a valley; that a glacier blocked the mouth of an ice free valley or that the proglacial drainage was impaired considerably.

**5.5.7 Erosional Forms**

The distribution of erosional features is clearly related to areas where bedrock is exposed. Striae are preferentially located on corrie lips, upside of protuberances, and in steep sided valleys at low altitudes, where bedrock is streamlined in direction of bedrock. Much of the bedrock in the field area is not ideal for striae preservation and this is reflected in the difference in the number of striae observed compared with Wester Ross. Roches moutonées and ice moulded bedrock occur in the larger valleys trending E-W. They are numerous in Glen Shiel and Glen Strathfarrar and are closely associated with areas of ice moulded bedrock and striated surfaces. Strongly striated surfaces indicate that the Loch Lomond Readvance glaciers were wet based, the distribution of ice moulded bedrock is probably controlled by the steeper slopes where velocities are inferred to have been greatest (Section 5.6). especially in the SE of the area.

**5.5.8. Variations in drift thickness**

In addition to hummocky moraine, glacial deposits occur in the form of undulating drift or till sheets. The thickness of this ground moraine varies considerably, its thickness being revealed by postglacial streams and gullies cutting through it as shown in Plate 5.5. It is clear that in many instances the occurrence of thick drift is related to topography. Numerous situations occur where it has been caused by ice moving up a reverse slope, for example in Strathcroe, north of Dorusduain at NG 982240 and in Coire Fionnarach (NH 1644). Similarly, thick sequences of drift, often associated with an irregular moudny surface or morainic ridge occur at a number of confluences within the Loch Lomond Readvance icefield, for example in NH 0424. Occasions where ice becomes thick on crossing a col occur at NH 1844, for example. Several thick sequences also occur that are specific to one locality. For example, the valley between An Riabhachan and An Creachal Beag (870 m) contains especially thick drift, which is probably due to ice having emerged from the eastern end of the Monar basin being forced up the valley. The
thick drift in Golden Valley (NH 0147) is probably a result of it lying transverse to the main glacier flow. Deposition would have been facilitated by the reduction in basal ice velocity and subsequent retardation of the basal layers of ice.

The topographic situations in which thick drift occurs, namely ice flow up a reverse slope, crossing a col or valley or at valley confluences are similar to those situations reported by Thorp (1984, 1991) for the area to the south of the field area. However, other factors are also important. The amount and location of acquired sediment from the Late Devensian ice sheet is important but ill defined. The occurrence of thick drift in small valleys below high corries such as Gleann a' Chiolich may be due to an increased debris supply from the periglacially-exposed slopes of Sgurr nan Ceathreamhnan. The distribution of thick drift accumulations is shown in Figure 5.27.

**Figure 5.27**
Distribution of Thick Drift in southern Ross-shire
5.5.9. Meltwater channels

Meltwater channels are abundant in the field area, occurring in a variety of situations. Proglacial channels are common emerging from the former maximum snout position of many glaciers. The most impressive of these are at NG 975337 and NG 970372, south and north of Glen Ling respectively, and at NH 255495 in Gleann Chorrain. Proglacial meltwater channels related to snout positions during downwastage also occur such as that at NH 015265. Ice marginal channels are also evident for example, at Glean nam Fiadh, north of Glen Affric, and NH159560, SE of Achnasheen. Comprehensive suites of meltwater channels which are of probable proglacial origin also occur in areas of kame and kettle topography and in areas of hummocky moraine, for example in Glen Moriston (Sissons, 1977d), Glen Orrin (NH 265465) and Strathcarron (NH 030480).

5.5.10. Orientations of drift

Orientations in drift in the form of flutes and/or medial moraines were identified in numerous localities in the field area. These are closely associated with hummocky moraine (Gray and Coxon, 1991). Whereas in the past orientations in drift have been described as fluted moraine (Robinson, 1977; Hodgson, 1982, 1986), this classification can be refined (Bennett, 1991; Bennett and Boulton, 1993b). Thus the small orientated ridges identified in the field area that occur within Loch Lomond Readvance limits may be identified as lateral moraines (e.g. NH 002335), or medial moraines (e.g. NH 073452) occurring below rock spurs and at glacier intersections, rather than simply flutes or fluted moraine (e.g. Rose, 1989b). Their distribution is shown in Figure 5.28

5.5.11. Discussion

The mapped evidence suggests that there is considerable variation in the spatial distribution of glacial and periglacial landforms within the field area. Factors affecting this distribution include lithology, aspect, altitude and topography (valley connectivity). The distribution of landforms and sediments reflects lithological controls to some extent. The presence of thicker drift in the north is possibly due to lithology. Here Torridonian Sandstone and Lewisian Gneiss are more common. The importance of lithological control on subglacial processes such as erosion is well established (Sugden, 1978; Boulton, 1974,1979) and may be reflected in the large scale distribution of glacial sediment assemblages. The lack of major moraine ridges, for example, may be due to the highly resistant rock types and similar explanation may be invoked to explain the paucity of striae on the Lewisian Gneiss. The distribution of certain landforms can be related to glacial dynamics (see below).
5.6. Reconstruction of the Loch Lomond Readvance glaciers

Reconstruction of the former extent and surface morphology of a former ice mass can yield valuable information such as the estimation of the former equilibrium line altitude and provides a basis for palaeoclimatic inference (e.g. Sissons, 1974b, 1980b; Ballantyne, 1989). The evidence described in the previous sections was used to reconstruct part of the highland icefield formerly present in southern Ross-shire. Reconstruction of the form of the transection glacier complex and associated corrie glaciers was achieved by estimating glacier surface contours at 50 m vertical intervals following the procedures developed by Sissons (1974, 1977b), using the altitudes of delimiting evidence and orientation of ice-directional evidence. Locating the position of the contours on the icefield and associated outlet glaciers was enabled by the abundant evidence for former ice directions, provided by striae and ice-moulded bedrock, erratics and streamlined drift forms. The field evidence thus severely constrains the positioning of such contours, as
glacier surface contours should be drawn approximately normal to the former ice movement direction.

Some subjective assessments of the glacier boundaries were necessary, for example, where several glaciers have a single source area or where the limit had to be interpolated between the highest glacial and the lowest periglacial evidence. In corries with glacially plucked backwalls the determination of the height of the ice occupancy in the corrie was difficult; Sissons (1983) assumed that the ice surface was 30 m below the top of the backwall (Gray, 1982; p.127), whereas (Thorp, 1981b; Gray, 1982) suggested that in some corries the ice was continuous with ice on plateaux or summits above the backwall. In southern Ross-shire the partial occupation of a corrie is often documented by the occurrence of limited areas of ice moulded bedrock above which lies extensive frost shattered bedrock and talus. In a few cases, the limit is inferred to lie completely within a corrie owing to the lack of evidence or, in places because of a peat cover. The heights of glaciers terminating in tidewater was assumed to be between 30 and 60 m. Flint (1971).

5.6.2. Distribution of glaciers

Figures 5.29 - 5.32 (in folder) show that the overall pattern of glaciation in the field area was one of valley glaciers descending from ice dispersal centres, chiefly nourished by corrie glaciers or snow gathering re-entrants on steep hillsides. Although described as a large icefield, the pattern of glaciers in southern Ross-shire, which forms part of what may be termed the 'West Highland Icefield', is more accurately described as a large transection glacier complex. In this complex the Monar Basin, fed by numerous corrie glaciers, may, however, be considered a small icefield with five outlet glaciers and various overspill tongues. The glaciers extended up to 24 km from the source areas, and many glaciers appear to have spread out on lower ground where they became free of the constraint of the confining valley slopes. The valley glaciers were for the most part separate but there was some connectivity through cols especially in the south of the field area, for example at NH 098239. Numerous nunataks were present at the glacial maximum and a number of corrie glaciers existed independently.

5.6.2.3 The ice divide

The position of the ice shed is shown in Figure 5.25. It is clear that some degree of asymmetry is present in the icefield with glaciers extending further east from the ice divide than to the west. This together with the curve in the reconstructed ice shed has palaeoclimatic implications which will be considered later.
5.7. Palaeoglaciology

5.7.1. Introduction

The reconstructions shown in Figures 5.29 - 5.32 were employed to evaluate the dimensions and characteristics of the former transection glacier complex. This was accomplished, by first dividing the icefield into component glacier basins along former ice shed defined by contours and landforms indicative of former directions of ice movement. Areas were derived for each individual glacier or glacier basin. A grid of squares, with sides equivalent to 50 m (25 m for the corrie glaciers) was superimposed on to maps of the former ice extent, and the number of squares that lay entirely or predominantly within the glacier limits counted, thus determining the implied area.

Volumes of ice were subsequently calculated by measuring ice thickness (glacier surface minus land surface altitude) at regularly spaced points within each glacier, calculating the average thickness and multiplying by glacier area. 5 - 30 sample points were used for each contoured interval. Glacier surface altitude was recorded at the same 5 - 30 sampling points. This information was used to calculate mean aspect, and ice-surface gradients at these points (calculated from horizontal distance between contours) and were then averaged to obtain mean gradients. The error margins for the transection glacier complex as a whole are thus regarded to be very small since the reconstruction is based on a great amount of evidence.

The combined measurements of ice surface gradient and aspect at each sample point were used to calculate an insolation factor, which represents the potential influence of direct insolation on each glacier for the probable ablation season (May-September), taking into account the transmissivity of the atmosphere and the albedos of snow and ice (cf. Sissons and Sutherland, 1976; Ballantyne, 1989). The lowest altitude of each glacier was obtained from Ordnance Survey maps.

Finally, the former equilibrium line altitude (ELA) for each glacier at its maximum extent was calculated as the area weighted mean altitude of the glacier surface (cf. Sissons, 1974). This procedure assumes that ablation and accumulation rates are linearly related to ice-surface altitude.

The results of the measurements and calculations outlined above are summarised in Table 5.2. These reveal that the Loch Lomond Readvance ice in southern Ross-shire at its maximum occupied a total area of c. 791 km$^2$ (90% icefield; 5% corrie G 1; 2). In terms of former ice volume; a total of 129 km$^3$ of ice was represented by the reconstructed glaciers, 99% of which was contained in the icefield.
Table 5.2
Data for glaciers in southern Ross-shire

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5.7.2. Basal Stresses

Glacier surface and basal long profiles of 15 outlet glaciers were drawn parallel to inferred flow lines and averaged over distances of 2 km along flow lines. Basal shear stresses were derived from these data according to the equation:

\[ t = \rho g h \sin \alpha \]

where \( t \) = basal shear stress, \( \rho \) = acceleration due to gravity, \( h \) = ice thickness in metres, \( \alpha \) = surface slope.

The equation, derived by Paterson (1981) assumes that ice behaves as a perfect plastic. There are errors associated with the calculations because of the limitations of the field evidence. Thorp (1991) suggests that calculated basal shear stresses are most accurate in terminal zones where the glaciers are well delimited and least accurate for source areas, but even so errors are unlikely to exceed 15 KPa. Figure 5.33 shows the distribution of calculated shear stresses calculated at 2 km intervals.

Basal shear stresses range from 6 to 396 KPa. The variation of these calculated basal shear stresses enables inferences to be made relating to the palaeodynamics of the former reconstructed glaciers and thus allows comparisons with results from other areas. Figure 5.33 shows that in 7 of the 8 glaciers for which there are complete data, maximum values occur towards the snout of the reconstructed glaciers. There is a distinct increase in basal shear stress values between 3 - 8 km from the glacier snouts. The values between glaciers vary considerably however and there are a number of apparent anomalies.

The low values of basal shear stress in the upper and middle reaches of most of the reconstructed glaciers may indicate that the ice was warm based (Paterson, 1981). High values recorded near the snout of the Cannich Fhiodhaig and Strathfarrar glaciers may be due to steep ice gradients or bedrock roughness. Low basal shear stresses correspond with valleys in which smoothed bedrock surfaces predominate, for example Glen Affric.

Comparing these data with those published elsewhere, reported basal shear stresses across the Antarctic ice sheet (Cooper et al., 1982) and Svalbard (Dowdeswell, 1986) also attest to low values in the vicinity of ice sheds and relatively low values near the margins of the icefields or ice sheets. Thorp (1991) also describes a similar pattern for an icefield with thirteen outlet glaciers in the western Grampians, with maximum values in the last 5 km.
Figure 5.33
Basal shear stresses and glacier profiles for glaciers in southern-Ross-shire
5.7.3. Thermal Regime

Inferences regarding the thermal regime of the former Loch Lomond Glaciers may be based on bedrock characteristics and drift characteristics. The abundance of glacially scoured bedrock which suggests basal sliding attests to the fact that the Readvance glaciers were predominantly warm-based and at pressure melting point. In addition to this, it has been suggested that warm based ice typically produces an assemblage of push and dump moraines, outwash fans and glaciofluvial forms such as eskers (Boulton, 1986; Boulton and Eyles, 1979) and sedimentological characteristics may include till, sand and gravel. Conversely, characteristic ice marginal landforms produced by cold based ice include large shear moraines, and hummocky bouldery topography (Weertman, 1961; Fitzsimmons, 1990). The landforms and sediments in southern Ross-shire identified above are consistent with an interpretation of warm based ice there during the Loch Lomond Stadial. The abundance of ice-marginal features both at maximum and recessional positions also attests to the former presence of warm based ice (Boulton and Eyles, 1979; Chorley et al. 1984).

5.8. Characteristics of the icefield and Palaeoclimatic reconstruction

5.8.1.

The characteristics of the ice masses and its glaciological attributes such as thermal regime can be evaluated from the reconstructions of the Loch Lomond Readvance glaciers. These data may be used to evaluate the impact of glaciers on the landscape and to determine some aspects of the palaeoclimate during the Late Quaternary (e.g. Manley, 1959; Sissons, 1974b, 1979a, 1980a; Sissons and Sutherland, 1976; Sutherland, 1984; Ballantyne, 1989).

5.8.2. Equilibrium line altitudes

It has been demonstrated in studies of the mass balance of modern glaciers that the equilibrium line altitude (ELA) provides a crucial link between glaciers and climate and is thus of great importance in palaeoclimatic investigations of glaciers (Sutherland, 1984). The ELA represents the altitude on a glacier surface where the exact balance between accumulation and ablation is achieved. If the ELA rises on a given glacier whose terminus does not move; it implies either an increase in snowfall or a decrease in the rate of ablation. It is suggested therefore, that the pattern of ELAs across an area may be interpreted in terms of palaeoclimatic controls on glacier form and distribution. In this study ELAs have been calculated for nine of the eleven valley glaciers since the remaining two have substantial parts lying to the south of the study area.
The pattern of calculated ELAs which rise from 481m to 670m is shown in Figure 5.34. With the exception of the ELA for Glen Affric, the ELAs depict a consistent trend for the transection glacier complex. The diagram illustrates that ELAs are higher for the eastern glaciers than those in the west of the area, and that ELA values also generally decline towards the North. The simplest explanation of the trend of higher ELAs in the east is that despite the greater build up of ice in the east, snowfall was heavier in the west allowing ice to build up in low lying and small accumulation areas.

**Figure 5.34**

Pattern of ELAs across the southern Ross-shire

The analysis of ELAs to the west of the study area by Robinson (1977) showed lower ELAs on south facing glaciers and she suggested that the dominant climatic conditions responsible for this were southerly snow bearing winds. In the SW Grampians Sissons (1979a) suggested a pattern of snowfall associated with S to SE air streams preceding fronts to explain the north-westward rising ELAs for that area. The study area is accessible to west coast influences and the pattern of ELAs reported here support the concept of dominant west to south-west snow bearing winds. This decrease in ELAs to the north is consistent with that reported for small corrie glaciers on the northern mainland (Sissons, 1977).
It is possible to compare the pattern of ELAs with those for other regions, some of which are depicted with trend surface analyses on published maps (Sissons, 1980b, 1983c). These maps demonstrate that during the Loch Lomond Stadial ELAs rose markedly away from the west coast into the highlands northwards. However, the data presented in the present study indicate that the interpolated lines are drawn too far west by varying degrees. Sissons (1983c) inferred that the primary control on differential glacierisation was the pattern of precipitation and he implied that the western coastal mountains were especially suitable for glacierisation, an assertion confirmed elsewhere (Cornish, 1981). To the south of Ross-shire Sissons and Sutherland also demonstrated a close association between the pattern of ELAs and the 'Highland Edge' - a topographic barrier exerting a considerable influence on precipitation in the SE and associated with depressions originating in the Atlantic, but this was not an influence in the study area.

5.8.3. Palaeoprecipitation patterns

From the above evidence it is clear that there was a rise in the snowline from the east to the west, and, in view of the cold climate, this must have been a function of variations in precipitation. At the present day rainfall varies from 2500 mm to 3000 mm per annum in the principal watershed area in Glen Shiel, decreasing eastwards to c. 2100 mm at the eastern end of Loch Cluanie, 1800 mm at the Mullardoch Dam and 1100 to 1300 mm near Dundregg gan and Tomich (NSHEB Annual reports 1956-66). It would thus appear that, during the period of the Loch Lomond Readvance, the proposed eastward decline in precipitation in the area as a whole was similar to that at present. Accordingly the palaeoclimatic inferences are consistent with Sissons and Sutherland (1976) interpretation of a stormy climate with frequent depressions tracking over northern Britain. They further suggested that this was a consequence of an oceanic polar front off the west coast of Britain generating and directing depressions across Scotland; an assertion supported by Ruddiman et al. (1977) and Duplessey et al. (1981).

5.8.4. Palaeotemperature implications

The pattern of ELAs together with estimated values for accumulation at the ELA during the Loch Lomond Stadial may be employed to estimate former summer temperatures. The relationship between mean summer (ablation season) temperature (t) to accumulation at the equilibrium line (which approximates accumulation(a) over the whole glacier), has been described by Sutherland (1984). Sutherland derived a regression equation of the form:

\[ A = 0.915 \cdot 0.339t \quad (r^2 = 0.989, P<0.0001) \]
(A is expressed in metres water equivalent and t is in °C.)

from data for ten Norwegian glaciers (c.f. Figure 4, Sutherland, 1984).

The glaciers considered by Sutherland (1984) occupy a marine location in Norway and it is considered reasonable to assume that the relationship between mean summer temperatures and accumulation at the ELA of the glaciers in southern Ross-shire took a similar form. In view of this, it follows that if the precipitation in southern Ross-shire during the Loch Lomond Stadial can be estimated, then the implied mean ablation season temperature can be calculated.

The impact of local controls such as variations in received insolation, aspect in relation to snow bearing winds and increased snow accumulation due to snow blowing and avalanching on regional temperature and precipitation has been noted in previous studies (e.g. Ballantyne, 1989). However, such influences are more important for small valley or corrie glaciers and are considered to be of minimal importance in the southern Ross-shire due to the fact that the icefield and associated glaciers were relatively large compared to potential source areas of snow blowing and avalanching. The contributing corrie glaciers for each valley glacier also tended to vary widely in aspect.

There is insufficient data available to compute temperatures for individual glaciers so these were grouped. Given that there are marked precipitation contrasts across the field area and that the area can readily be divided into eastern and western glaciers, separate calculations from the respective ELAs were computed in addition to a regional one. The area weighted mean ELA for the western flowing glaciers is c. 570 m. That for the glaciers descending along the eastern valleys is c. 580 m. The mean annual precipitation for the west at 570 m is 2750 mm and for the east at 580 m is 2200 mm (Meteorological Office, 1977). The inferred precipitation totals for the western area are 2500 - 3000 mm and for the east 2100 - 2300 mm. It has been suggested, however, that it could be expected that 20 - 25% of this would have occurred as rain in summer (cf. Manley, 1959; Sissons and Sutherland, 1976). Taking this into account, the inferred mean annual snow accumulation is between 1.6 and 1.8 m for the W and for the E 1.9 - 2.4 m water equivalent. These values may be substituted into the regression equation given above. A range of possible mean ablation season (May - September) temperatures at the ELA may be derived if accumulation is assumed to be 70% of present day annual precipitation (i.e. the percentage of the annual total that falls in the winter months), i.e. about 1750 - 2101 mm in the west and 1470 - 1610 mm in the east at the respective mean ELAs. It should be noted that this is derived from an ELA associated with a situation of equilibrium when temperatures are likely to have ameliorated slightly (Ballantyne, 1989); mean July temperatures during the stadial may have in fact been slightly lower than this. Data from previous studies have related accumulation at the ELA to summer temperature thereby
allowing temperatures to be inferred. These are 2.1 -2.3 °C in the west and 1.9 - 2.0° C in the east. Assuming a lapse rate of 0.6°C/100 m, this gives temperatures of between 5.0° C - 5.2° C and 5.6° C- 5.7° C for the east and west respectively.

The calculated palaeotemperatures can be compared with those from other areas. The mean sea level temperature for southern Skye, calculated by Ballantyne (1989) is 6.3 +/- 0.5° C which is considered a low estimate for that area. That for Wester Ross is 6° C (Ballantyne et al. 1987) Calculated palaeotemperatures for the Loch Lomond Stadial appear to increase southwards with mean July sea-level temperatures of 7.8° C calculated for the western Grampians (Sissons, 1979a, 1980a, 1980b) and 7.5° C for the Lake District. The figures for southern Ross-shire are consistent with data for nearby areas. The data for this area, however, incomplete and the derived figures can only be regarded as estimates at present.

Certain problems can be identified concerning the above procedures. The approach employed assumes a steady state equilibrium between the climatic system and that of the ice mass. The validity of such an assumption during what have been described as periods of rapid climatic change has been questioned (e.g. Bennett, 1991; Bennett and Boulton, 1993) in terms of lack of contemporaneity of the maximal limits. For the southern Ross-shire ice-field as a whole, however, the occurrence of ice dammed lakes between glaciers at maximal limits suggests that limits, at least in close proximity, achieved maximal limits contemporaneously and the individual glaciers considered here are sufficiently discrete to yield results for individual glaciers as well as the icefield as a whole.

5.9. Deglaciation of the Loch Lomond Readvance icefield

Both biostratigraphic and geomorphological evidence have been employed to attempt to determine the nature of ice decay and pattern of deglaciation for various parts of Scotland (Eyles, 1983; Sutherland, 1984b; Thorp, 1984, 1986,1991; Walker and Lowe, 1985; Benn et al. 1992). Prior to the beginning of this study, however, there had been no published information regarding the pattern of deglaciation in southern Ross-shire, except for that undertaken by Charlesworth (1955). Just prior to completion of this study two publications (Bennett and Boulton, 1993a, 1993b) appeared, which addressed this topic in the wider context of the Scottish Highland icefield Their data and conclusions will be evaluated in relation to the present fieldwork.

The nature of deglaciation is clearly defined in a number of locations, which are described below. From a maximal position identified in Figure 5.10, the nature of the retreat of glacier ice in Glen Carron is recorded by abundant evidence. An extensive area of hummocky moraine extends from Loch Gowan (NH 1556) to approximately 3 km west of Loch Sgamhain around NH 075515. The configuration of much of this moraine
suggests that the ice retreated actively, westwards from the limit at Achnasheen. Major recessional or possibly readvance positions, identified by descending drift limits, aligned hummocky moraine and fragmented ridges are located on the south side of the valley e.g. NH 1554 to NH 1453, with associated ice marginal positions in the adjacent valley for example at NH 157535. Later standstills during retreat also occur at NH 122537 and NH 100529 in Glen Carron. Evidence in the corries to the SW of Achnasheen shows that at an early stage of deglaciation, one and possibly two small corrie glaciers retreated into the two western corries, whilst the lateral margin of the main valley glacier deposited a moraine, now fragmented below the corrie descending from 490m to 450 m ((NH 1251). There is also evidence in the form of fragmented shoreline terraces that an ice dammed lake in Glen Carron was formed during this period of deglaciation. The lake at its maximum extent probably extended from NH 077518 to NH 127542 at a level of about 200 m (Figure 5.35). Further west in the vicinity of the ice shed, sub parallel

Figure 5.35

Ice-marginal features in Glen Carron
benches occur on the south side of Ciolle Bhan at 0651 and around Meall an Fhluichaird NH 070495. These occur between c. 430 and 260 m, and probably record pauses in the downwastage of the ice here, as less ice flowed into the main valley from the south and from the easternmost Applecross hills. The lake described above probably ceased to exist as lobes of ice retreated actively up the Lair and Fionn Abhainn valleys to the North, and south across Meall an Fhluichaird and Coire ant Seilach into Pollan Buidhe. Successive lateral moraines also attest to a gradually diminishing mass of ice in the upper part of Strathcarron (Figure 5.32). The continuation of active retreat back into the corries is indicated by the presence of moraines in the corries to the north of Sgurr a Chaorachain and Bidean an Eoin Deirg, and Coire Leiridh and Coire na Eilde to the north of Sgurr na Feartaig.

Clear recessional moraines with associated laterals are apparent at NH 175490 and NH 160485 in the valley of the River Meig. (Figure 5.36). Moraines at the mouths of corries such as Fuar Tholl Mor and east of Maoile Lunndaidh also indicate that active retreat prevailed until deglaciation was almost complete here. It appears, however, that following the prolonged stage of active retreat, in situ stagnation may have occurred with chaotic mounds being deposited within such corries.

**Figure 5.36**
Recessional moraines in Glen Fhiodhaig

The eastern outlet glaciers terminating in Glen Orrin and Glen Strathfarrar received ice from the Monar Basin. Ice also flowed out of it to the west into Glen Ling. Whilst distinct retreat moraines are present in the main valleys of these eastern valleys, the pattern of retreat to the west appears more complex. The topography between the Monar Basin and Glen Orrin suggest that ice supply to Glen Orrin was terminated by falling ice surface levels and that stagnation occurred at a relatively early stage in the main part of
The valley. The Strathfarrar glacier appears to have retreated actively, receding downslope rapidly rather than upvalley. This would also appear to be the case for the Monar basin in general, a situation indicated by the successive lateral moraines especially notable on the south-eastern slopes of Maole Lunndaidh and Carn nam Fiaclan for example at NH 145435 (Figure 5.37).

Figure 5.37
Recessional moraines at NH 145435

Cross-valley ice marginal landforms at NH 270335 and NH 203318 provide limited evidence for glacial retreat stages in Glen Cannich. However, there are substantial lateral moraines at numerous locations attesting to successively lower levels of the former ice surface, and stillstands. In Glen Affric several major recessional positions are evident at
NH 277282, NH 249262, NH 185230 and NH 130205. These limits are associated with the successive laterals found at NH 236270 and NH 277220. Glen Elchaig also supports recessional moraines occurring at relatively wide intervals up the valley. This may suggest a number of pulses of rapid retreat with distinct pauses resulting in moraine formation. The frequency of ridges increases towards the head of the valley.

Only one major recessional position occurs in Glen Shiel, this is located in the valley in which Loch Duich occurs, at NH 901225, as indicated by valley side evidence and submarine investigations. In the tributary valleys of these glens, however, numerous recessional stages may be identified such as those in the valley of the River Glennan and in Coire Dhuinnid.

The major retreat stages identified in the present study are distinguishable from the numerous recessional ridges that occur close to these larger moraines or the more chaotic hummocky material between them. The pattern or spacing of retreat positions identified in each valley is of some importance, with regard to the sequence of deglaciation in the respective valleys. The distances between major recessional ridges are shown in Figure 5.38 together with information describing the presence or absence of a moraine in the contributing corries. It is noted that there is no consistent pattern with regard to the relative size of moraines, although the outermost one is usually largest.

**Figure 5.38**

Pattern of retreat stages across southern Ross-shire
In only a few cases are multiple ridges present at the maximum snout position, for instance in Glen Cannich and Glen Orrin. The spacing of features relating to recessional stages is fairly regular in many valleys. In Glen Carron, suites of moraines occur at 1.5 km, 4.25 km and 8.5 km upvalley from the former glacier snout. In Glen Fhiodaig the distances are 3.25 km, 5 km and 7 km. Recessional stages in Glen Cannich occur at 2 and 8.5 km and in Glen Affric they are located between 5 and 7 km, at 10 km and 12 km from the inferred glacial downvalley maximum. It has been noted that the pattern of retreat stages is more complex in Glens Orrin, and Ling. In Glen Orrin, *in situ* stagnation appears to have been important whilst in Glen Ling numerous recessional stages are identified. It is also noted that in the upper portion of many of the main valleys, the density of recessional ridges and associated landforms is greater.

Eleven distinct valley glaciers exhibit evidence for a maximum of four major periods of interruption of deglaciation, probably relating to periods of climatic oscillation (see below). This is not unexpected in view of the behaviour of similar recent and present day glaciers in Norway and the multiple termini of Loch Lomond Readvance glaciers for many other locations in Scotland (e.g. Loch Treig, Sissons, 1967).

The main conclusions that can be drawn from the evidence for the nature of deglaciation in southern Ross-shire is that there was not a rapid and immediate amelioration of climate in the late stadial, either in terms of temperature or precipitation, but fluctuations in one or both of these parameters caused certain glaciers to respond with successive halts or in some cases readvances during retreat. The wider significance of these conclusions is considered below.

The extent of the Loch Lomond Readvance in southern Ross-shire and the pattern of deglaciation as reported in this thesis is at variance with the data and conclusions of Bennett and Boulton (1993a, 1993b). This is exemplified in Figure 5.39 where the present work is shown together with Bennett's (1991) reconstruction. These differences reflect, in particular, different strategies for the interpretation of drift deposits, especially hummocky moraine. Bennett and Boulton assume that hummocky moraine, following Horsfield (1983), is the product of active ice and that alignments of ridges relate to the dynamics of the ice and can be interpreted as representing former ice margins. However, investigators such as Lowe and Walker, (1981), Thorp, (1984, 1991), Walker and Lowe, (1985) and Tipping, (1988) have interpreted hummocky moraine as a product of ice stagnation and thus it has no marginal significance This debate is not resolved and it has been suggested that both forms of hummocky moraine develop (Benn, 1990 and Benn et al. 1992).
In Bennett's (1991) reconstruction marginal positions are effectively traced outwards from the ice shed area as a series of concentric and increasingly older margins. Limits in adjacent valleys are linked by assuming similar patterns of ice retreat in each valley. In his doctoral thesis, Bennett in fact states that 'for the most part the limits presented are simply interpolated between moraine fragments'. In addition to this, Bennett described difficulties in defining a limit due to the 'proliferation of possible limits' and the chosen limits may therefore be rather arbitrary.

It is not surprising given the assumptions made in their reconstructions that the defined Loch Lomond Readvance outer limit in this study differs from that of Bennett (1991) and Bennett and Boulton (1993a) in almost all of the mapped valleys. The maximal limits in Glen Orrin, Glen Cannich, Glen Ling and Glen Affric are more extensive than the present study by 10 km. Their suggested positions coincides in places with outwash terraces mapped in this study. Conversely, the limit in Glen Strathfarrar proposed here is further downvalley than that of Bennett and this is also the case for Glen Elchaig, Loch Duich by c. 10 and 4 km respectively. The only area of agreement with the present study, is at Achnasheen where the eastern limit for the Loch Lomond Readvance is formed by
a continuous moraine limit that can be traced laterally over 30 km, inside of which there are no moraines as distinctive or laterally continuous.

Further, ice thicknesses determined in this study on the basis of trimline evidence including lateral moraines is very different from that of Bennett and Boulton (1993). In their model for ice thickness Bennett and Boulton (1993) project a surface profile from the terminal areas towards the ice shed; as a consequence, their ice mass is over a hundred metres lower than that described for this study.

The current mapping, based on more detailed evidence, does provide a more accurate limit for the Loch Lomond Readvance in southern Ross-shire than that just described. The limit between valleys is elucidated in the present study and the eastern and western limits reconciled. The trimline evidence also provides a means of determining the true lateral extent of Loch Lomond Readvance ice. Bennett and Boulton do stress, however, that their research is not a detailed reconstruction of glacial history but directed towards gaining an understanding of glacial dynamics and response to climatic change.

In addition to differences concerning limits, this study suggests that the pattern of deglaciation was not a progressive and continuous withdrawal of ice to the ice shed. The pattern of deglaciation indicated by the presence of numerous recessional moraines indicates that the climate did not simply ameliorate in response to a simple rise in temperature or decrease in precipitation. It is more likely that fluctuations in either or both of these resulted in successive stillstands during the process of retreat.

5.10 Conclusions
The mapped evidence has established both the lateral and vertical extent of Loch Lomond Readvance ice and enabled the form of a transection glacier complex in southern Ross-shire to be reconstructed which consists of eleven major outlet glaciers together with corrie glaciers. The pattern of former ice flow from a central N-S ice shed is identified. The characteristics of glacial and periglacial landforms in the area is also described. It is difficult to determine the degree of erosion that is attributable to Loch Lomond Readvance glaciers, especially with regard to corrie formation. In relation to the volume of debris in moraines and fluvioglacial deposits, however, the implication is that a relatively small percentage of the material removed from most corries is attributable to the Loch Lomond Readvance glaciers. It is suggested that the last glaciers modified a landscape primarily eroded in previous glacial events and this in turn would imply that a similar pattern of ice build up in corries expanding into valleys occurred previously, and almost certainly earlier in the Late Devensian.
From the mapped evidence and calculated glacier dimensions, various inferences regarding palaeoclimate and former glacial dynamics have been outlined which are comparable to those of similar studies conducted in other areas of Scotland.

In contrast to the recent work undertaken by Bennett and Boulton (1993), work largely based on air photo mapping and limited field evidence and not intended as a 'definitive statement', the present research defines a maximum limit for the lateral extent of the Loch Lomond Readvance based on detailed field evidence. Although Bennett and Boulton (1993) state that 'given that the icefield may have been over 50 km wide (Peacock, 1970), an error of a few kilometres in locating the limit is unlikely to affect the overall pattern', the present author suggests that the maximal limit proposed by Bennett and Boulton is markedly over or under - estimated in places. This also accounts for the differences in the projected ice thicknesses compared with those in the present study.
CHAPTER SIX

The Late Devensian Ice-sheet in Southern Ross-shire.

6.1 Introduction

The present chapter considers evidence for a glacial event which pre-dates the Loch Lomond Readvance and is presumed to be Late Devensian in age. Little published information is available with regard to ice-sheet glaciation in southern Ross-shire. Much of our existing knowledge dates to research carried out by officers of the Geological Survey almost eighty years ago. In the light of recent detailed studies in other areas of Scotland (e.g. Thorp, 1987), which have added to knowledge of ice sheet characteristics and advanced our understanding of the dimensions, patterns of movement, ice sheet source and divide areas and glaciological aspects of the last ice sheet, this study attempts to apply similar detailed analytical study to southern Ross-shire. The following sections are devoted first to describing the inferred pattern of ice movement associated with the ice-sheet, second to determining the maximum extent of this ice sheet and third to assessing the glaciological implications of the available evidence.

6.2 The last ice-sheet: patterns of ice movement.

6.2.1 The nature of the evidence.

Erosional evidence such as striae, ice-moulded and glacially-sculptured and streamlined bedrock forms, friction cracks, crescentic gouges and roches moutonnees are present in the field area above and beyond the mapped limits of the transection glacier complex. Together with streamlined drift and erratics these features are presumed to relate to the Late Devensian ice sheet.

Figure 6.1 depicts all known striae, friction cracks and strongly streamlined bedrock forms outside the defined local glaciation given in 5.29-5.32. The distribution of such ice-directional evidence is clearly sporadic. In the south of the field area it tends to be sparse because the later icefield reached an elevation of over 800 m and obliterated evidence from the earlier ice sheet. The centralmost area (around 10°E) is also relatively devoid of such features for the same reason. The effects of periglacial frost shattering is a second factor influencing the distribution of erosional ice-direction indicators: striae on mountain tops are more likely to have been destroyed by the effects of weathering than those below c. 600 m. The distribution of ice-moulded rock in particular is also determined to some extent by altitude. It occurs primarily below c. 500 - 600 m, and where it is present above this, tends to have been at least partially obscured due to the effects of periglacial weathering. Further, drift is fairly widespread over the study area.
and obscures ice directional features. Although it is possible for drift to become streamlined beneath the ice sheet, little evidence of this nature was found in southern Ross-shire. Exceptions occur in Glen Ling and Glen Orrin where limited fluted depositional forms occur in a downvalley direction.

**Figure 6.1**
The distribution of striae outside the Loch Lomond Readvance Limits

Lithology would appear to assert some degree of control over the occurrence of ice-directional evidence. The mapped distribution of striae suggests that these features are preferentially preserved on the Cambrian Quartzite in the north east and on the granulitic and siliceous schists rather than the pelitic gneiss of the Moine Series. The Lewisian
Gneiss also appears to support many striae. Finally, striae within the limits of the later stage are found to have been crossed by other striae that are clearly related to ice movements of Loch Lomond Readvance glaciers. Some of these are believed to the product of the Late Devensian ice sheet, having survived the later glaciation, but they have been employed with circumspection.

Although almost all of the above-mentioned forms of glacial evidence are located outside the limits of the last phase of glaciation, and predate this phase, the question remains as to whether they are of Late Devensian age or indeed predate the ice sheet and are remnants of an even earlier glacial event. Their age is equivocal but it may be argued that it is unlikely that small scale erosional forms at least, would survive the intense periglacial conditions that may have existed on any nunataks. It is also doubtful whether such forms could have been preserved under the renewed glacial activity of the Late Devensian and subsequent periglacial conditions of the Loch Lomond Readvance. Similarly, it is likely that any glacial deposits laid down by pre-Late Devensian ice sheets were removed by the last ice sheet. It is probable that the Late Devensian ice sheet developed on a landscape that had characteristics already created by the erosion of previous ice sheets. For this reason, the use of the alignment of glacial troughs and breaches as ice-directional indicators is considered to be of limited value.

Many studies have used the distribution of erratics to establish flow lines in order to determine the patterns of ice-flow direction. The homogeneous lithological character of much of the study area restricts the usefulness of this technique for establishing ice flow patterns. Where lithological differentiation does permit identification of erratic trains in the field area, this is in almost all cases within the Loch Lomond Readvance limits and therefore probably relates to ice movements during the later glaciation. With regard to ice sheet movements, erratics of Moine Schist from within the field area have been identified to the north on the Torridon Sandstone and Cambrian Quartzite of Wester Ross, for example. Peach et al (1913a) report fragments of Moine schist on the south-eastern slopes of Sgorr Ruadh up to a level of 830 m, on An Ruadh-stac and to the north beyond these hills. This makes it possible to extrapolate flow lines from the furthest extent of a train of Moine Schist back into the field area, although the widespread scatter of such erratics and large potential area of origin is such that it is possible to conclude only that the ice carrying these boulders moved in a westerly or north-westerly direction across Glen Carron.

The possibility of establishing further erratic trains was examined in relation to the limited areas of Ross-shire where the rock type is sufficiently lithologically distinct to enable former ice flow reconstruction. A number of igneous intrusions notably the Cluanie granodiorite pluton were identified and surveys undertaken on ground adjacent to these in an attempt to locate erratic trains issuing from them. Erratic boulders were
evident to the north of this outcrop above the maximum altitude of the Loch Lomond Readvance glaciers, extending in a north-easterly direction, confirming an observation to this effect by Peacock (1971). Granodiorite boulders are absent from the slopes of Sgurr nan Conbhairan to the West and its summit area which lends support to the evidence for a former eastward ice sheet flow pattern.

The relative value of the varying forms of evidence depends first on its type and second on its location both in terms of altitude and topographic position and its location within the field area as a whole. Evidence located in a mountain-top situation provides a minimum altitude for the extent of the ice sheet at its maximum and is likely to relate to ice-maximum movements. Evidence in valleys, however, may relate to post-maximum readvances or to movement during downwastage. It is necessary to analyse striae in relation to adjacent topography: topography may be responsible for generating the general pattern over the area as a whole or may create local distortions. It is useful to consider each piece of evidence in terms of assessing the location of such data in relation to altitude and the orientation of the (pre-existing) glacial troughs, as well as viewing the pattern as a whole.

6.2.2. Inferred nature of ice movement and ice accumulation areas.

The position of the ice shed is important in order to ascertain the distribution of source areas and characteristics of the Late Devensian palaeoclimate. This is shown in Figure 6.2, together with the flow lines inferred from Striae (Figure 6.1) and other ice directional indicators. The defined ice shed relates to the highest striae observed in the area but it is possible that they may not relate to the actual ice sheet maximum but to a post-maximum position. However, there is no evidence for possible movements of the ice shed and the postulated divide is presumed to relate fairly closely to the maximum extent of the Late Devensian ice sheet.

On higher ground ice directional indicators shows a fairly simple North-South divide from which ice generally flowed to the NNW and NNE. This is supported by striae observations both in the north of the area (e.g. Beinn Tharsuinn NH055433 and Carn Gorm NH136500) and in the south of the study area (e.g. Beinn Fhada NH018193 and Aonach Shasuinn NH168182).

The ice shed zone is defined by the westernmost limit of indicators for east-flowing ice and the eastern limit of evidence for the western movement of ice. In places this represents a narrow zone less than 1 km wide whereas in the centre of the field area this zone is up to 5 km wide. It is not possible to define the location of the ice divide within this zone. The northerly component identified in the ice flow pattern is strongest towards the south of the field area and may reflect flow patterns away from more
Figure 6.2
Patterns of ice movement of the Late Devensian ice-sheet in southern Ross-shire
mountainous areas or 'topographic highs' in the southern parts of the field area. The data also suggests that there was a substantial northerly flow originating in an area to the south of southern Ross-shire. This northwards movement of ice contradicts the alignment of most of the principal valleys which tend to be orientated east-west, and overrides them. Evidence for this is found on both summits and intervening ridges and confirms that such a movement was probably associated with the ice sheet maximum as such high level data are more likely to represent the true direction of flow at ice sheet maximum unrestricted by topographic control.

A secondary trend in ice sheet ice flow directions occurs in the form of a radial pattern, evident in the vicinity of the Monar Basin, although the striae supporting this form of movement occur at a lower elevation than those striae indicating the general pattern described above. In Upper Glen Orrin, ice passed over Bac An Eich in NNE direction and NW of Creag Coire na Feola it flowed to the NE; a northward movement is registered on Garbh Carn and between Carn Eitige and Sgurr Fearstaig. Other striae reveal a southwards movement away from the Monar region: in Glen Orrin deflections occur to the SE; on Beinn Mheadhoin to the ESE and on Carn Na Cre to the ENE. The Strathfarrar hills show ice movement in a ENE to NE direction. These directions are consistent with both diverging and radial flow (the Monar Basin itself being oriented W - E). The fact that the striae indicating radial flow contradict the orientations of nearby striae at higher altitudes confirms that this flow is likely to be related to a post maximum movement. Accordingly, this evidence supports the situation of ice spilling out of the basin (not over the mountains) when the ice sheet declined.

A third pattern of strong valley alignment in the lower parts of many of the valleys is also indicated in the striae data. Glen Carron, for example, cuts across the NW of the field area in a south westerly direction and many ice directional forms on the valley sides and adjacent hills are orientated downvalley, such as those on Carn Mor. This also appears to be the case in Glen Attadale. There is also a strong downvalley element (to the SW) in the upper Glen Ling region superimposed on the general NW trend; for example, striae and roches moutonées show a WSW trend on Beinn Dronaig and to the SW on Carn na sean Luibe.

In addition to these general patterns, isolated deflections, unrelated to valley constraint, also occur. In lower Glen Orrin for example, evidence for a diverging ice stream is apparent where striae are found to diverge in the lower part of the valley.

6.2.3 Discussion

The patterns of flow described above have implications for assessing the nature of accumulation patterns in the field area. It suggests that there may be a strong
topographic control of ice flow. The inferred patterns of ice movement and ice divides are broadly coincident with previous studies although some reinterpretation of the evidence is possible.

Previous studies have advocated the presence of a North - South ice shed, extending through the NW Highlands from Sutherland to the Cowal Peninsula (Figure 6.3). Peach et al. (1913a) on the basis of a more limited number of high level striae placed the ice shed considerably to the east of the West Monar mountains. In the Rannoch Moor area, Thorp (1987) found that the pattern of ice domes was controlled by the distribution of mountain massifs. The influence of the mountain massifs in the field area appears to have been minimal in terms of their effect on ice flow across the area. It is likely however that the Carn Eige massif and the west Monar hills were major sources of ice dispersal although evidence for their importance is only found in terms of their post-maximum influence on ice flow.

Figure 6.3
Ice movement associated with the last ice-sheet, after Price, 1982
The interpretation of high level striae presented here accords with the proposed 'last ice sheet' (stage I) of the Geological survey (Peach et al. 1913a). The stage III glaciation proposed by the survey is less extensive than that demonstrated for the Loch Lomond Stadial in this study and accordingly does not fully support the earlier interpretation of striae evidence. The most significant deviation from former interpretations is the assertion in this study, as elsewhere in Scotland (e.g. Thorp 1987), that the data demonstrating topographically constrained ice movement relate to the downwasting of the ice sheet below the summits in a progressive manner and not to a separate 'confluent glaciers' stage.

The evidence of striae at different elevations in the region of the Monar Basin suggests that the pattern of radial flow was a post maximum dispersal centre in agreement with the suggestions of the Geological survey (Peach et al. 1913a). This evidence is quite compelling, but the data might also be interpreted as being due to either the presence of an ice dome that was contemporaneous with the N-S ice shed, or the deflection of ice by topography under the ice. There is however, little evidence to support the idea of an ice dome at the ice maximum as there are few indicative high level striae and no reason for a deflection of the ice shed to the East over the Monar Basin. The second idea of deflection in this manner during the maximum does not generally correspond with the topographical constraints of the localities in which such deflections are evident.

It is clear that changing ice flow directions did occur during the decay of the last ice sheet but the data do not unequivocally support glacier resurgence during decay. Although minor deflections occur throughout the field area these may be attributed largely to the influence of topography, for example, where basal ice was deflected around a positive relief feature. The presence of such a pattern of changing flow directions, however, does advocate a process of active retreat. The evidence may, therefore, be explained in terms of changing flow patterns with an uncomplicated north-south divide at the Late Devensian ice maximum followed by the active retreat of the ice. A possible readvance of large valley glaciers at some stage, perhaps filling the Monar Basin and overflowing in many directions through cols in the constraining ridges, cannot be excluded on the available evidence.

The position of the ice shed, the patterns of ice flow and inferred accumulation areas may be compared with those in adjacent areas (Figure 6.4). These are broadly in agreement with the position given in this study. To the North in Wester Ross striae and the distribution of erratics show that the pattern of ice flow was similarly characterised by a westward to north-westerly movement across all but the most easterly parts of the area. Here, however, there is strong evidence for ice shed movement, although the timings of such movement is unknown. In particular, eastward and westward migration
of the ice shed is necessary to explain the north-westerly and easterly carry of Inchbae augen - gneiss from its outcrop (Ballantyne et al., 1987).

Figure 6.4

Ice movements and ice divide for the Rannoch Moor Area (Thorpe, 1987)

Recorded evidence in the area between Skye and the field area suggest a deflection of mainland ice from a south westerly direction to a north-westerly direction. This is due to the presence of a separate ice centre over Skye. Its influence appears not to have extended to affect the westward flowing ice within the field area since the westerly ice sheet flow patterns of the eastern part of the field area are mirrored beyond its eastern boundaries (Peach et al., 1913a). The area immediately to the south of southern Ross-shire has been neglected with regard to studies of glacial history. Features of former ice sheet movement in western Inverness-shire are only discussed briefly by Peacock (1970a), and flow lines depict ice movement towards the west from an area to the west of Loch Eil, Loch Arkaig and Loch Quoich. Peacock suggests however, that the ice shed lay well to the north-east of Loch Eil (and east of that deduced for the Loch Lomond Readvance).
Although many of the earlier researchers advocated the eastern positioning of the ice sheet divide relative to the Loch Lomond Readvance ice shed, this view has been challenged. In the western Grampians the ice shed (and associated ice domes) is now believed to have been situated over Ben Nevis, somewhat to the west of that previously postulated. Recent studies, such as that of Thorp (1984, 1987) have suggested that the distribution of Loch Lomond glaciers provides an analogue for the build up of the last ice sheet and that the position of the ice divide closely approximates that of the Loch Lomond Readvance. The present study confirms that the ice divide in southern Ross-shire approximated that of the Loch Lomond Readvance and was further west than the position suggested by Peach et al (1913a).

6.3 The Dimensions of the Late Devensian ice sheet.

6.3.1. The lateral extent

Chapter 2 described continuing controversy with recent shifts in opinion regarding the extent of the last ice sheet: from the traditional view in which the ice sheet was believed to be confluent with the Scandinavian ice sheet (e.g. Boulton et al 1977; Anderson, 1981), consensus changed to support the view of an ice sheet of very much more limited extent with parts of Scotland ice free (Sutherland, 1984). The latest suggestions continue to hold that the lateral margins of the last Scottish ice sheet extended beyond the outline of the present coastline terminating at the Wee Bankie Formation and Bosie Beds in the North Sea (Sutherland, 1984; Bowen et al. 1986; Serjup et al. 1987; Nesje and Serjup, 1988) and into the Minches off the west coast of Scotland (Bowen et al. 1986; Sutherland, 1984) with an independent ice cap co-existing in the Outer Hebrides and evidence for a marine embayment. Despite the fact that the ice sheet is now believed to have been of more limited lateral extent, the whole of the field area must lie within the lateral margins of the former ice sheet at its maximum extent. Previous research to the east of the field area has recorded former ice marginal positions in the lower portions of the valleys of the Orrin and Conon beyond the easternmost boundary of the field area, and has demonstrated that the extent of ice was beyond the Beauly and Moray Firths. Several successive ice marginal positions (e.g. Firth, 1984, 1989) relating to ice sheet deglaciation have been documented there. Southern Ross-shire also lies within the western lateral margins identified by Peacock (1980) and by Ballantyne (1989). The lateral extent of the ice sheet clearly has implications for its vertical dimensions.

6.3.2 The vertical extent of the last ice sheet.

6.3.2.1. The suggestion that the last ice sheet may have been of comparatively limited extent implies that many mountain summits may have risen above the last ice sheet as nunataks
and may have remained exposed to intense periglacial weathering above the surface of the last ice sheet throughout the Late Devensian or longer (Boulton et al., 1991).

Evidence in support of this suggestion has been found in parts of northern mainland Scotland and on Skye where high level periglacial trimlines were identified in Wester Ross and in Trotternish. These data were interpreted by Ballantyne (1991) and Reed (1988) as representing the former surface of the last ice sheet: above the 'trimline' is thick Late Devensian (or pre-Late Devensian) periglacial mountain top detritus and below it the bedrock retains elements of ice moulding. (Figures 6.5 -7 demonstrates the nature of the trimlines previously identified in northern Scotland.) The conclusion was drawn that the uppermost parts of some Scottish mountains remained ice free as nunataks at the Late Devensian maximum. This work however, remains controversial and these are the only areas for which ice sheet periglacial trimlines have been reported.

Figure 6.5
Trimlines identified in Wester Ross

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**Legend**

- ▲: Trimline separating in situ mountain top detritus from ice scoured bedrock
- ▲: Mountain summits supporting in situ mountain top detritus and mass removed detritus
- □: Mountain summits supporting incipient mountain top detritus
- □: Ice scoured mountain summits
Figure 6.6
Trimlines identified in Trotternish, Skye (Ballantyne, 1989)

Figure 6.7
Area of western Grampians wholly covered by ice (Thorpe, 1987)
In the Rannoch moor area to the south of the study area, Thorp (1987) demonstrated, that the ice sheet surface was above all mountains including Ben Nevis (1344m). The present field area lies between Ben Nevis and Wester Ross and therefore southern Ross-shire is in a critical position for establishing whether the pattern of trimlines in Wester Ross is continued farther south and if the ice sheet gradient is apparent and is constant. If the gradient described for northern Ross-shire and for Trotternish is correct the higher mountains in southern Ross-shire would also have been nunataks and support corroborative trimline evidence.

Past research has invoked the existence of cold-based ice (Sugden, 1966, 1978) to explain variations in mountain top detritus and ice moulding. This possibility must also be considered, together with the various other factors which may contribute to changes in bedrock and detritus on mountains. Further, Gordon (1979) mapped areas of areal scouring in relation to hypothetical ice sheet surfaces over southern Ross-shire. Thus, the field data collected on trimlines can be used to assess the accuracy of the above concepts as well as addressing the question of whether the ice sheet was the most limited extent proposed (e.g. Sutherland, 1984) or whether the slightly more extensive cover of ice (Hall and Bent, 1991) appears more likely.

6.3.2.2 Analytical methods

The vertical extent of the ice sheet over southern Ross-shire was investigated by examining high level summit and spur characteristics outside the inferred Loch Lomond Readvance limits. This was carried out in order to first, determine the character and distribution of mountain top detritus and ice scouring characteristics, and second to ascertain the possible constraints on the altitude of the ice sheet. The methodology of identifying and mapping mountain top detritus and bedrock features has been outlined in Chapter 4. As well as identifying the distribution and characteristics of bedrock and detritus, attempts were made to locate and identify trimlines, to establish the lower limit to in situ mountain top detritus and to assess the distribution of these in relation to lithology, structure, slope and altitude.

The trimline concept and related 'nunatak hypothesis'

Research on ice sheet trimlines in Scotland has apparently been stimulated by first, the concept of a Scottish ice sheet of limited extent and second, by work undertaken in Canada on the eastern margins of the Laurentide ice sheet and in Norway on the western margins of the Scandinavian ice sheet. In Scotland the presence of 'mature' mountain top detritus has been shown to define areas that were not glaciated during the last glacial maximum (cf. Ives, 1978). In Scotland widespread areas of in situ frost weathered detritus have been identified which have been shown to be Late Devensian in age and to be relict (Ballantyne, 1984). Trimlines have also been observed since the last century (Forbes, 1846; Harker, 1899, 1901).
However, there are several problems associated with the technique of using trimlines and with the 'nunatak' hypothesis. The technique is based on defining the downslope limit to \textit{in situ} mountain top detritus and the upslope limit to glacially scoured bedrock. There are difficulties with the identification and use of trimlines: a \textit{distinct} and abrupt downslope limit is reported in some areas of northern Ross-shire, in other areas, the altitude of the surface has been interpreted from less distinct 'trimzones'. The clarity of the transition depends amongst other factors on the local effectiveness of glacial abrasion and the extent to which frost weathered debris moved downslope following glacier downwastage. In his doctoral thesis, Reed (1988) suggested that the ice sheet surface tended to be represented by a less obvious 'trimzone' occurring over tens of metres. He describes trimzones occurring over altitudinal ranges exceeding 100 m, for example on Sgurr nan clach Geala in the Fannichs. These concessions may invalidate certain assumptions made regarding the close relationship of the trimline to the ice sheet surface.

Second, a fundamental difficulty concerns the age of the periglacially weathered detritus occurring on mountain tops. Important criteria for identifying the upper, older trimline is the thickness or maturity of \textit{in situ} frost weathered detritus. These are often difficult to assess but are important criteria for identifying this upper (older) trimline. A third problem concerns the fact that bedrock surfaces are likely to be differentially weathered due to variables such as altitude, lithology, structure, slope gradient and former exposure to ice or periglacial weathering (Ballantyne, 1984). Other factors such as dip of strata may also be important. With regard to lithology, it has been demonstrated by Reed (1988) that trimlines are preferentially preserved on certain lithologies and poorly developed on others. In particular, it was noted that trimlines or zones were poorly preserved on Moine Schists, which underlie most of the field area. Accordingly, trimlines, if present, are likely to be poorly developed and this required that all changes in the thickness of mountain top detritus and the nature of the bedrock had to be carefully mapped and analysed.

Features important for the elucidation of possible trimlines include the range, size and type of periglacial landforms, the depth and type of mountain top detritus, bedrock characteristics and their location relative to summit areas and slopes. Information was thus collected as follows: mountain top detritus was classified with regard to its thickness and type (in situ or mass-moved; I, II, or III (Figure 2.9; and thickness on a scale, very thick >1m to incipient <0.5m). In places where bedrock cropped out at the surface it was assessed for the degree of frost shattering (determinable by the angularity, amount of evident macro- and micro-gelivaton and depth of open joints). Details of lithology, structure, dip, aspect and slope were also noted.
The higher ground within the field area where pre-Late Devensian features are likely to have been preserved can be divided into nine main massifs: 1. Achnasheallach hills 2. Strathfarrar 3. The Killilan hills 4. West Monar Hills 5. An Riabhachan-Sgurr na Lapaich 6. Carn Eige massif 7. Beinn Fadha-A’Glasbheinn 8. North Glen Shiel ridge 9. South Glen Shiel ridge. These are shown on Figure 6.8, with their main summits. Figure 6.9 comprises extracts from 1:25000 field maps that show the distribution of mountain top detritus and periglacial features for the major massifs in southern Ross-shire. Table 6.1 describes characteristics of mountain summits and independent hilltops, and cols in terms of periglacial detritus, geology, bedrock characteristics and provides points which may enable maximum and minimum altitudes for the last ice sheet to be determined.

**Figure 6.8**
Major mountain massifs in southern Ross-shire
Figure 6.9
Mountain top detritus and periglacial features on a) Maoile Lunndaidh
b) The Strathfarrar Hills

Key on p. 68 (Figure 4.1).
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Table 6.1

Characteristics of mountain summits in southern Ross-shire
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A' Chralaig INH094(48)
An Riabhachan NE (NH13934R)
Tom .' Choinich CNH163273)
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Saurr na Camach (NG977159)
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6.3.2.3 Mountain top characteristics in southern Ross-shire

The evidence for the altitude of the ice sheet is discussed for each mountain group with reference to the variables described above. The distribution of mountain top detritus and associated trimlines and bedrock characteristics is considered in terms of the 9 defined areas described above.

1. The Achnasheallach hills are located furthest to the north in the field area and, therefore, may be expected to be potential nunataks due to their proximity to identified nunataks to the north. There is, however no unequivocal upper trimline present. The mountain top detritus on the summit area of Moruisg appears very thick but is less so on Sgurr nan Ceannaichean. A lower limit to this thick mountain top detritus can be discerned between 860 and 870 m on Moruisg (Plate 6.1). The lower limit to mountain top detritus varies between 730 and 825 m on Sgurr nan Ceannaichean, although it is less thick there. The ice moulding retained on the col between these mountains demonstrates that the ice sheet surface was above 740m. There has been a limited amount of periglacial mass movement on these summits with soliflucted lobes and boulder lobes and sheets at 910 m. Soliflucted drift occurs on the southern slopes to 540-560 m. The summit areas of Carn Gorm (875m) and Carn Liath (857m) support thick mountain top detritus. To the west the summit at 627m (NH163517) displays incipient mountain top detritus and retains evidence of ice moulding to the NE.

Plate 6.1

Deflation surface and thick mountain top detritus, Moruig 920m
2. The Strathfarrar Hills  On these hills it is only on the summits of Sgurr na Fearstaig and Sgurr Fhuar-thuill that there is thick (>0.5 m) mountain top detritus. A distinct downslope limit to this is evident at 930m on Sgurr na Fearstaig and at 960m on Sgurr Fhuar-thuill and may represent the upper limit of the ice sheet. The other mountains in this range have mountain top detritus that extends as debris mantled slopes to a relatively low altitude to the Loch Lomond Readvance limits in many localities. The cover appears to be very regular - it maintains its thickness to these lower levels with considerable evidence for mass movement and supports a wide range of periglacial features including large solifluction lobes and sheets with numerous ploughing boulders. On Creag Ghorm a' Bhealaich there is however a change with mountain top detritus suddenly at 825 m becoming incipient (<0.5 m) and a similar change at between 860 and 875 m on Carn Nan Gobhar, on Sgurr na Ruidhe at 830-840m. The presence of a blockfield at 950m on Carn an Gobhar suggests prolonged severe periglacial conditions. There is little evidence for ice - scouring here as most of the hillslopes have been strongly periglacially weathered. Meall Dubh and Druim Dubh are ice moulded to their summits.

3. The Killilan hills support no thick mountain top detritus and therefore no upper trimlines. The summits of Ben Killilan at 753m, Sguman Coinntich 879m and Faochaig 868m are relatively low and covered with a thin veneer of incipient mountain top detritus. The summit and upper slopes show distinct signs of ice moulding to the west and north west and the ice sheet surface can be placed above c 800m here (Plate 6.2 and 6.3). Small boulder lobes and sheets occur on these hills and must therefore have developed since the ice sheet maximum. The summit of Cnoc na Uan exhibits distinct roches moutonées.

4. The west Monar Hills exceed 1000 m and may be expected to have formed nunataks during the Late Devensian ice sheet maximum. There is no widespread occurrence of a trimline although there are instances where the distribution of detritus may be interpreted as representing the upper surface of the ice sheet. A distinct downslope limit to thick mountain top detritus occurs on Maoile Lunndaidh (Plate 6.4) at between 920m and 950m, below which the detritus is much thinner. Similar changes occur at altitudes of 950-960 m on Carn nam Fiaclan, and 870-890m on Sgurr Coinnich. The thickness of in situ mountain top detritus on Bidean an Eoin Dearg and on Sgurr a Chaoarachain is variable although some change is evident at 960m and 880m respectively and many of these slopes are steep and have mass moved detritus. Bidean a choire Sheasgaich is too steep to support thick debris mantle and the top is ice moulded implying that the ice sheet was > 940 m here. A gradual change in the thickness of the mountain top detritus occurs on most of these hills down to the level of the Loch Lomond Readvance limits, and the lower summits support incipient mountain top detritus and frost shattered-ice moulded bedrock. Carn Guerdain (594m) and Ben Dronaig
Plate 6.2
Ice scoured summit of Sgumann Coinntich

Plate 6.3
Slightly shattered ice scoured gneiss, 700m on Sgumann Coinntich
exhibit ice moulded bedrock that has been slightly periglacially modified. Eagan (690m) is well ice moulded. Conversely Beinn Tharsuinn is very frost shattered but with no thick weathered mantle: detritus is patchy but where it occurs supports solifluction and boulder lobes. Sgurr na Feartaig is well ice moulded with patches of mountain top detritus.

Plate 6.4
Incipient mountain top detritus on Maiole Lunndiadh, 950m

5. An Riabhachan-Sgurr na Lapaich
No distinct changes have been identified as the cover of mountain top detritus is very variable. There is evidence for a change in detritus thickness on Carn nan Gobhar where there is thick mountain top detritus to between 975 and 920m with a blockfield between 975 and 950m. The summit of Creag Dubh at 940m supports thick detritus. On An Riabhachan a change in drift thickness occurs at between 780 and 880 m with mass moved mountain top detritus to 740m. Sgurr na Lapaich supports thick mountain top detritus to 990m. Elsewhere the depth of this periglacial detritus appears less than on the mountains described above to the north.

6. Carn Eige massif Possible trimlines occur on the Carn Eige and Sgurr nan Ceathreamnan summits where in situ detritus is thick to an altitude between 1050-1110 m and down to between 960-1030m respectively. This is present only on the west-facing slopes in both cases which are not cut into by corries. Elsewhere such different weathering zones are less apparent and the lower limit of thick mountain top detritus can be placed at 790-830m on Beinn Fionnladh, 950-970m on Mam Sodhail, 940-950m on
Tom a Chionnich and 975m on Toll Creagach and 990m on Sgurr na Lapaich. The mountain top detritus on Mullach na Dherigain and An Socach is thinner. An Soutar (676m) and surrounding summits on this plateau to the east of the Carn Eige massif are all ice scoured with little modification.

7. Beinn Fadha-A Glasbheinn-Sgurr an Airgid range
The highest of these adjacent mountains in the SW of the area, Ben Attow exhibits a distinct change in the depth of weathered detritus at 975m. The surrounding summits of A' Glasbheinn (918m), Carnan Cruithead (730m), Beinn Bhuidhe (702m), Carlan nan Euan (594m) and Sgurr an Airgid (841m) have ice moulding up to their summits and show signs of only very limited modification by periglacial weathering.

8. North Glen Shiel ridge comprises numerous steep spurs and summit areas that tend to be ice moulded and supporting little in situ mountain top detritus. The summits of Sgurr a Bhealaich Deirg (1031m) and Sgurr an Fhurail (988m) and Am Bathach are clearly well frost-shattered but support only patchy detritus in which are small turf banked terraces. There is thick mountain top detritus evident to c 980m on A' Chralaig and 960-990m on Mullach Fraoch-choire. Aonach Shasuinn (889) has a cover of incipient mountain top detritus with small solifluction feature but retains numerous ice sheet moulded outcrops that protrude above the debris. Similarly the top of Tigh mor na Seilge is well ice scoured and retains the form of a large roche moutonée. It supports only patchy incipient mountain top detritus. Sgurr na Conbhairean is the only mountain where there is a possible occurrence of a trimline, as thick mountain top detritus extends down to 1050m, below which there is ice scouring. The plateau area of which Carn Ghlusaid is a principal summit is covered by a thin cover of superficial frost weathered detritus. The outcrops that occur are frost-shattered but preserve evidence of ice sheet scouring; the ice sheet was therefore > 957 m here.

9. The South Glen Shiel ridge is similarly steep and devoid of trimlines. The cover of mountain top detritus is variable. There is evidence for ice moulding by the ice sheet up to 900m.

6.3.2.4. Assessment of evidence for ice sheet reconstruction
A major aim of the investigation into the Late Devensian glaciation of Southern Ross-shire was the identification of a possible upper limit of the ice sheet following the methodology of Reed (1988) and Ballantyne (1991). However in the field area the data do not replicate the evidence found to the north in Wester Ross. The data were first inspected in order to identify locations where thick in situ mountain top detritus was separated by a sharp break from ice moulded bedrock in such a way as to clearly identify the presence of an upper trimline above the Loch Lomond Readvance limits. Such a distinct line as described by Ballantyne (1991) was apparent nowhere within the region. Many of the summits and upper slopes in the field area support at least 'incipient' mountain top detritus and the distribution of ice moulded bedrock is sporadic. The absence of clear trimlines on most of the highest mountains means that there is no firm
evidence for establishing the actual altitude of the ice sheet surface. Where the distribution of mountain top detritus can be interpreted in terms of an indistinct trimline, for example on Maoile Lunndaidh and Ben Attow, the assertion that this may have been an ice sheet nunatak is not corroborated by the evidence on adjacent mountains and there is, therefore, no evidence of this kind for an ice sheet surface increasing in altitude towards the south. There is no reason to suppose that these mountains had more favourable conditions for trimline formation as the rock type does not vary significantly and adjacent hills are significantly lower.

Reed (1988) and Ballantyne et al. (1987) have no consensus on the description of mountain top detritus occurring above their upper trimlines, referring to it sometimes as 'mountain top detritus' or on occasion 'thick mountain top detritus'. Although mountain top detritus is widespread, the concept of a trimzone as defined by Reed (1988) is restrictive for the description of field relationships observed in the field area. In southern Ross-shire data indicate that the transition from the mountain tops to ice moulded bedrock or mass moved mountain top detritus elevations is best described as a weathering zone. Some of the debris mantled surfaces of this type are a product of in situ weathering of bedrock but in other areas, especially on slopes and lower plateaux, detritus is clearly glacigenic and has often been modified periglacially. It was noted that there were difficulties in defining the precise transition from in situ mountain top detritus to mass-moved mountain top detritus, but abrupt changes in the thickness of mountain top detritus and associated periglacial features do occur. (Figure 6.10). However, given that the thickness of such detritus is variable in southern Ross-shire and the tendency for Moine Schist to have weathered at least to some extent during the Loch Lomond Stadial (Reed, 1988), then it was considered probable that only very thick (>1m) detritus is likely to represent areas not covered by the last ice sheet.

On several mountains there was a great altitudinal difference between the upper limit of clear ice moulded bedrock and the lower limit of in situ mountain top detritus. It is possible in such cases that a maximum altitude for an ice surface could be ascertained from the lower limit of mass-moved mountain top detritus, given probable known rates of movement for certain types of mass movement such as solifluction lobes (Ballantyne, 1984). This would not apply to debris that had been moved by rapid mass-movements or in the presence of steep slopes.

In the field area many summits showed evidence of ice scouring and were clearly below the surface of the Late Devensian ice sheet. It was thus possible to establish minimum altitudes for the ice sheet surface using such summit data points that were clearly scoured by the last ice sheet.
**Figure 6.10**

Lower limit to mountain top detritus

![Map showing lower limit to mountain top detritus](image)

**Figure 6.11**

Upper limit to ice-moulded bed-rock showing minimum altitudes for ice-sheet ice

![Map showing upper limit to ice-moulded bed-rock](image)
6.3.2.5. The ice sheet surface in southern Ross-shire
Positive statements may be made about ice sheet elevations on the basis of the field observation, but the data cannot be explained by a simple model of the ice sheet surface as proposed by Reed (1988) for Wester Ross.

1) The only possible trimlines identified in the field occur on Maoile Lunndaidh at 950 m and Ben Attow at 975 m.

2) There does appear to be a general rise in the altitude of the lower limit to in situ mountain top detritus towards the south. Accordingly ice sheet altitudes rose from 800-960 m in the North to 920-1090 in the Monar region to 810-1020 and then 970-1060 in the south of the field area. The distribution of the lower limit to mountain top detritus is too irregular to confidently confirm the occurrence of a series of nunataks.

3) A greater proportion of summits in the south retain elements of ice moulding. Minimal altitudes of glacial scouring can be determined across the area. These are shown in Figure 6.11. Well preserved ice moulding on many of the Glen Shiel Hills suggests that the ice was above c.1000 m there and over 1000 m at places in the Carn Eighe massif. Similarly, the bedrock at localities such as Bidean Choire a Sheasgaich and Sgurr na Muice imply that the ice sheet passed over 900 and 850 m respectively, although a very limited amount of periglacial modification has ensued. The fact that ice moulding is evident over 900 m at locations throughout the field area implies that the whole of the field area was covered by the last ice sheet.

Clearly, these conclusions are not compatible with a simple ice sheet surface rising to the south as might be expected if the trend reported to the north continued into this field area. The data are examined in more detail below and alternative models of the ice sheet explored.

6.3.2.6. A Lateglacial origin for mountain top detritus?
Before investigating alternative ice sheet models to explain the nature of mountain top detritus in the study area, it is necessary to establish that the mountain top detritus present is too thick to have developed since the ice sheet maximum. Several researchers (e.g. Crampton and Carruthers, 1914, Galloway, 1961 and Peach et al., 1913) interpreted the mountain top detritus as being a product of intense periglacial weathering during the period of downwastage as the summits became exposed or during the renewed cold conditions of the Lateglacial period. It has been suggested more recently that the thickness and character of the detritus in places is such that it is unlikely to have formed in the last 18 000 years (e.g. Ballantyne, 1984; Reed, 1988) and debate has focused on the means by which it has been preserved.
The existence of high level ice moulding is a fairly strong indicator that the ice sheet passed over bedrock surfaces above 1100 m in the south and at least 900 m in the north of the field area. The ubiquitous occurrence of frost shattering implies that very few surfaces have been left intact due to periglacial weathering during downwastage and during the Lateglacial. The fact that the detritus on many of the mountains is produced from both micro- and macro-gelivation implies that it is pre Late Devensian in age and may have formed in the period between the onset of the last glacial cycle and its maximum. Although micro-gelivation has been prevalent during the Holocene, it is likely that the considerable amount of macro-gelivation was produce over a much longer time than that of the Lateglacial.

A number of alternative explanations for the distribution of ice moulded bedrock and mountain top detritus in southern Ross-shire may be offered in terms of particular field relationships. These are discussed below.

**Ice moulding**

A greater proportion of summits in the south retain elements of ice moulding. There appears to be a clear relationship between the amount of ice moulding with altitude for individual mountains. Locations likely to have been under the ice sheet, if high, tend to be frost shattered. Many hills retain some evidence of having been glacially scoured whilst supporting a thin veneer of frost-weathered debris. There appears to be a gradual change in the degree of frost shattering with altitude.

**The Lower limit of thick mountain top detritus**

Mountain top detritus occurs on the majority of mountain summits outside the Loch Lomond Readvance limits. Its thickness varies considerably and changes in this thickness are in some places abrupt and others gradual. On some mountains the mantle completely blankets the surface and bedrock is absent; on others the debris cover is interrupted by step-like rock outcrops with debris infilling the treads between them. Such differences probably reflect the resistance to weathering of the underlying rock.

The distribution of thick mountain top detritus appears to reflect slope angles on the mountains: there will be no thick mountain top detritus if the slope is too steep and this may explain the lack of thick mountain top detritus on many of the Cluanie Hills.

There is thick mountain top detritus on mountain summits at altitudes of less than 800m. It is very unlikely that such summits could have been exposed as nunataks at the ice sheet maximum. Given established facts about the ice sheet surface and flow patterns, it might be expected that the lower limit of thick mountain top detritus would be similar on adjacent mountains or comply with the constraints of an ice sheet surface. In fact, the distribution of thick mountain top detritus also changes abruptly across summit areas.
independent of altitudinal changes, especially on plateaux areas. Over the field there is evidence for large scale periglacial mass movement of mountain top detritus and for widespread downslope movement of much of the thick mountain top detritus.

6.2.3.7. Local controls on the distribution of mountain top detritus

It is clear that there is little consistency between mountains in terms of its altitudinal distribution. The distribution must therefore be controlled by another factor or a number of other factors combined. Factors that may influence this distribution include altitude, lithology, structure, slope.

(i) Altitude.

The fact that there are few instances in which there is a transition from ice moulded bedrock to thick mountain top detritus over a few tens of metres, would appear to support the hypothesis that the distribution is a function primarily of altitude. The depth or thickness of the mountain top detritus often changes very gradually with altitude. The changes of mountain top characteristics with altitude are recorded on Figure 6.12. If altitude were the single controlling factor on the distribution of mountain top detritus (and the whole area was under the ice sheet), then it would be expected that there would be a gradual change with the highest summits exhibiting the thickest mountain top detritus. This is not the case (see below). On some individual mountains, however, this pattern does seem to be sustained.

**Figure 6.12**

The change of character of mountain top detritus with altitude
(ii) Lithology.

The presence of trimlines is clearly affected to some extent by the lithology. As Reed (1988) demonstrated, they are generally absent on the Moine schists (which underlies most of the field area) but are distinct on other lithologies such as Torridon Sandstone. They are also present in Wester Ross on the Lewisian Gneiss which also occurs in the field area. In southern Ross-shire, the Moine metamorphic rocks and Lewisian Gneiss vary enormously in composition and structure and consequently in their response to periglacial weathering. Mica schists and pelitic schists tend to be weathered by processes of macrogelivation by means of extensive fracture along cleavage planes to produce platy clasts of variable size and marked angularity and the micaceous rocks are more susceptible to microgelivation. It is, therefore, likely that the mica-schists and related lithologies would readily weather to form debris mantled slopes comprising a diamicton of angular clasts embedded in a sandy-silty matrix. The more massive schists and gneisses would be weathered to form more equidimensional clasts and microgelivation is likely to have been less intense. These differences are reflected in the distributions in southern Ross-shire.

A number of areas of blockfield have been identified in this study, the largest of which occurs on Carn nan Gobhar in the Strathfarrar hills. Although it has been suggested that the presence of blockfields attests to more intense periglacial conditions in the past, it is more probable that the presence of a blockfield here relates to a local occurrence of a particular rock type or to a localised change in the rock structure. In the field area there is little appreciable difference in the amount and thickness of mountain top detritus between the different lithologies. With regard to the occurrence of ice moulding, the Lewisian Gneiss appears to retain features produced by glacial scouring more readily.

(iii) Structure.

Reed (1988) suggested that it is the jointing differences that are responsible for degree of susceptibility to frost action. In the field area there is no contrast such as that in northern Ross-shire regarding the rock types with markedly differing jointing patterns. This study has indicated that the dip may be just as important. Little attention has been paid to the dip of the strata in relation to surface topography and to the presence of faults and the effect of their presence on the nature of the mountain top detritus. It is important to consider the angle that the strata makes with the surface topography as this determines the proportion of planes of weakness in the bedrock that are exposed to atmospheric processes and therefore susceptible to weathering (Figure 6.13).

(iv) Slope. The slope clearly affects the distribution of detritus on upper slopes. Little detritus will be preserved on mountain tops if the slope is too great (cf. Ballantyne, 1984) and therefore slope determines the distribution of summits with in situ mountain top detritus across the field area.
6.4. Glaciological considerations - a cold based ice sheet?

A possible explanation of the pattern of ice moulding and mountain top detritus observed in the field area is that *in situ* mountain top detritus on some areas of high ground may have been preserved under relatively thin cold based ice whilst lower slopes were scoured by warm based ice (e.g. Sugden, 1968, 1977; Holdsworth and Bull, 1970; Boulton, 1979; Watts, 1983; Ramussen, 1984). The possibility of differential preservation of detritus and ice scouring is difficult to refute. Mature blockfields have been revealed by the shrinkage of thin cold based ice in Northern Norway (Whalley *et al* 1981). Similarly, Sugden and Watts (1992) have provided convincing evidence that tors and blockfields on Baffin Island have survived the passage of the last ice sheet. This has also been demonstrated to have occurred in the Cairngorms (Sugden, 1968).

The evidence in the field area supports this hypothesis, especially in the north of the field area where there is abundant thick mountain top detritus at high altitudes and evidence for ice moulding on lower slopes. In the south, the evidence for glacial scouring is much more widespread. Selective glacial scouring clearly occurs in the field area to high altitudes and is not simply related to relative altitude. The mapped distribution of thick mountain top detritus is largely coincident with the distribution of plateau areas in the field area (e.g. the plateau areas of the Monar Hills and the Strathfarrar hills). The topography of these contrasts with the ridges and peaks that are more prevalent in the
south of the field area, where there is greater evidence for ice scouring on mountain summits. The idea that plateau areas promote the occurrence of cold based ice and that ridges are associated with warm based ice moving between ridges and therefore ice scouring has been previously reported Sugden (1966). The question arises as to whether or not cold and warm based ice can coexist in close proximity.

It is possible that topography controls the occurrence of cold and warm based ice. The greater thickness of ice in the valleys may have produced warm ice. This may also be related to ice flow patterns: the ridges in the south of the area had large ice streams moving between them; flow patterns suggest that the plateau areas in the north and east may have had a cover of thin cold based ice. It is also possible that these plateau areas acted as accumulation centres and were the centres of initiation of flow, where the velocity of ice was probably relatively low (Nye, 1991) whilst the ice in the valleys was thicker, consequently warm based and therefore of greater velocity. Alternatively, it is possible that the ridges and peaks in the south only support thick mountain detritus where it has been able to accumulate readily. This argument is less convincing because there are usually some areas of flat or gently sloping surfaces at points on these mountains where debris could have accumulated.

There are several principal criticisms against this argument. First, the often abrupt nature of the boundary between glacially abraded and frost weathered terrain conflicts with the expected diffuse pattern, given the tendency of the boundary between erosive warm based ice to passive cold based ice to migrate during ice sheet build up and decay, although there will always be an upper limit to warm-based ice. Secondly, a regular and consistent down-glacier descent in trimline altitude is indicative of a 'true' periglacial trimline that marks the maximum altitude achieved by the former ice surface. Thirdly, the weathering limit is often continued downvalley with ice marginal depositional landforms such as lateral moraines (Grant, 1977; Nesje et al. 1987, 1988). These trends are not apparent in southern Ross-shire in relation to an 'upper' trimline, and the nunatak hypothesis is not upheld by such arguments.

Partial confirmation that a cold based ice sheet existed in the area comes from the work of Gordon (1979). In this work he examined the relationship of mapped geomorphological evidence to predicted basal thermal regime and ice sheet dimensions for two transects across Scotland, one of which passed through the study area. He tested the hypothesis that there is a spatial correlation between zones of reconstructed basal ice melting and landscapes of 'areal scouring'. Although 'areal scouring' refers to ice scoured bedrock and 'knock and lochan' topography, it may also include areas of ice-moulded then frost-weathered bedrock and is thus a much broader concept than ice moulded bedrock as used in this study. Thus Gordon's and this study are not fully comparable. The model also only tests data on a small scale, although it does examine
the evidence over a wide area. This approach may thus mask local factors producing changes in thermal regime, in which case local changes in thermal regime may be superimposed on a more general regional pattern of basal ice characteristics.

The concept of 'areal scouring' relates to long term evolution of the landscape possibly through several glacial cycles (Sugden and John, 1976). Some of the geomorphological evidence used by Gordon in southern Ross-Shire relates to the Loch Lomond Stage glaciation, so care needs to be taken in applying his model where the mapped ice scouring is located within the limits of the Loch Lomond Readvance and is not primarily resultant from ice sheet basal ice characteristics.

The pattern of areal scouring in two zones away from the central region is reflected in areas immediately to the north of Gordon's transect within the present field area. In Gordon's model, basal freezing is predicted for the central area. The predicted area of basal freezing should be coincident with areas where there is a consistent cover of mountain top detritus (that has been preserved under the cold based ice). Current field data indeed reveal that most mountain summits and upper slopes in parts of the central area do support a thick detrital cover. It is certainly only in the central areas (predicted to have been under cold based ice) where thick or very thick mountain top detritus has been preserved under cold based ice. The pattern is less easily determinable than that for outer regions as the central area was subjected to extensive erosion in the course of the Loch Lomond Readvance. The fact that some evidence for ice scouring exists on selected summits may also suggest, however, that other factors contribute to the pattern of scouring/ basal freezing.

In addition to examining the relationship between areal scouring and thermal conditions for a full sized ice sheet, Gordon (1979) also investigated the thermal characteristics for a smaller half sized ice sheet. He noted that the predicted zones vary under different conditions so that an ice sheet of half height produces a similar arrangement of zones with a smaller central area of basal freezing. The mapped evidence corresponds more readily to an ice sheet of more limited extent. The evidence does not wholly correspond to the model, especially with regard to the predicted zone in the centre of the area where, although there is evidence that may be explained in terms of the presence of cold based ice, there is also evidence for ice scouring. An alternative or additional explanation for the existence of warm based ice that was responsible for areas of ice scouring is considered below.

6.5 Discussion and Conclusions

The combined evidence indicates that the last ice sheet covered the whole of the field area and its surface was at least at 900 m in the north and greater than 1100 m in the south.
The gradual transition from ice moulded bedrock to thick mountain top detritus implies that altitude is a significant controlling factor on the distribution of mountain top detritus and preservation of ice moulded bedrock. However, the evidence is inconsistent areally and factors such as the dip of the rocks are considered important in this respect.

Thick mountain top detritus occurs on plateaux areas and suggests that many mountain summits were covered by cold based ice. The presence of mountain top detritus on larger summit areas and its absence on summits adjacent to steeply sloping hillsides where warm-based ice is more likely to have existed may explain this distribution (see below). It is feasible that independent cold based ice caps covered and protected some debris mantled plateaux whilst the surrounding valleys were occupied by powerfully erosive ice streams. It has been shown that even deeply weathered gruss on Scottish mountains apparently survived successive glaciations under the protective cover of plateau ice domes (Hall and Mellor, 1988). This scenario seems less plausible on steep aretes that have debris cover, however. In southern Ross-shire there are few aretes where thick mountain top detritus is preserved and this may account for the lack of thick mountain top detritus in the south where there are fewer plateaux. An alternative possibility is that the thick mountain top detritus formed since the ice sheet maximum. This, however, is considered unlikely.

The distribution of mountain top detritus and ice scouring is consistent with probable variations in the thermal regime of the former ice sheet, which was controlled by both the ice sheet thickness and topography, the former influencing the wider distribution and the latter determining local changes in this distribution.
CHAPTER SEVEN

Late Glacial - Interglacial transition: biostratigraphic investigations and interpretations

7.1. Introduction
Palynological investigations have been employed to reveal a further aspect of the Late Quaternary environment of southern Ross-shire. The aims of the pollen analyses are four-fold: the biotic and minerogenic characteristics of the deposits provide information about the Late Devensian and Holocene environment in terms of vegetation development, palaeoclimate and contemporaneous geomorphological processes; palynological investigations also provide a means of substantiating the hypothesised Loch Lomond Stadial age of the transection glacier complex described in Chapter 6; the analysis of the nature of the early postglacial record and comparison of this record between sites also has potential for establishing the pattern of deglaciation of the Loch Lomond Readvance ice mass in southern Ross-shire. The study also offers possibilities of examining the value of pollen deterioration studies in the context of Loch Lomond Readvance deglaciation. A fourth aim is to establish whether southern Ross-shire developed a vegetation similar to that to the west, with its marked oceanic influence or whether it shows a closer resemblance to that to the east.

7.1.2 Establishing a Loch Lomond Readvance age for glacier limits
The techniques for investigating 'inside-outside' sites were first employed by Donner (1957) and have since been used widely (e.g. Walker, 1974; Lowe, 1977; Walker et al 1988) to establish a Younger Dryas age for corrie and valley glaciers. Such methods have assisted the delimitation of the Loch Lomond Readvance throughout Scotland. At present there are no established pollen sites in southern Ross-shire with which to compare glacial evidence. This study attempted to rectify this situation.

7.1.3 Pollen deterioration studies
At all sites, deteriorated pollen was recorded and evaluated with regard to its origin and palaeoenvironmental value. Results from research into deteriorated pollen and its relationship to pollen profiles and their interpretation suggest that there may be considerable errors in certain pollen techniques as practised to date (Tipping, 1988). Interpretation of early Flandrian pollen is undertaken with close attention to the state of deterioration of pollen grains.

7.1.4 Elucidating the nature of vegetation development between the Loch Lomond Stadial and the appearance of tree taxa in the Flandrian
The need for detailed analyses of the Lateglacial - Interglacial transition in Scotland has been highlighted by recent studies (Walker and Lowe, 1987, 1990). The reasons for
differences in the nature of vegetation colonisation between sites in the area and in adjacent areas are addressed.

7.1.5 Establishing a vegetation sequence related to deglaciation

The possibilities for relating the lateglacial and early postglacial vegetation history of regions glaciated during the Loch Lomond Readvance to the glacial evidence have been explored in recent years. It has been suggested that it is possible to reconcile palynological evidence with glaciogeomorphic evidence and that it is probably necessary to combine evidence from geomorphic mapping with bio- and lithostratigraphic evidence and pollen analysis to obtain an accurate picture of the lateglacial-early postglacial environment. There is often considerable uncertainty regarding the extent of glaciers and as to which of several limits in a valley represents a Loch Lomond Readvance maximum. Tipping (1984) attempted to resolve this for the region between Loch Etive and the Cowal Peninsula, by selecting a number of critical sites in relation to glacial limits already established in the area. Although his results were inconclusive, a similar approach is adopted for Glen Affric.

7.2 Pollen sites

7.2.1 Experimental coring and site selection

In order to establish a Loch Lomond Readvance limit in a chronostratigraphical framework, suitable sites were sought for pollen-stratigraphic investigation in the vicinity of a distinct end moraine. Attempts were made to locate a suitable site as close to the moraine as possible, both inside and outside the former limit and a number of possible sites were identified using aerial photographs and by mapping.

A programme of experimental coring was initially undertaken in Glen Elchaig, Strathconon and Glen Affric. In Strathcarron and Glen Carron, preliminary coring investigations had already been carried out (Sissons, 1982) and it was considered that Lateglacial sediments must be buried under deep outwash sands and gravels, if present at all. Numerous locations were cored in the search for a tripartite stratigraphy (i.e. minerogenic material between organic deposits, all of which overlie impenetrable minerogenic sediment or bedrock) which would imply a Lateglacial succession of deposition and associated plant macro-fossils. An unequivocal sequence such as this was not located. Three sites were nevertheless chosen, primarily on the basis of their relation to former Loch Lomond Readvance glaciers, for further investigation: GA III at Loch na Gobhlaig, situated within 3 km of a very clear end moraine in Glen Affric, and a further two sites, 2.1 km and 12 km respectively inside the limit, for the purposes of comparison in terms of the vegetation changes relating to deglaciation.
7.2.2 Site location and description

7.2.2.1. Glen Affric I
This site is nearest the source area of glacier ice, located at NH 173228, 12 km from the limit of the northern lobe of the Glen Affric glacier, terminating at NH 298285 and 11 km from site GAI (Figure 7.1). It occurs at an altitude of 240 - 250 m. The dominant species in the present vegetation are *Molinia caerulea* and *Sphagnum* spp. *Calluna vulgaris* is also present together with *Erica tetralix* in places. *Potentilla erecta*, *Pinguicula vulgaris* and *Carex* were also observed.

7.2.2.2. Glen Affric II (Geusachan)
Despite extensive investigative coring upvalley of the Loch Lomond Readvance limit, no sites were found that contained an appreciable thickness of sediment. In the absence of a deep basin, a shallow core was obtained from a small ill-defined basin, fairly close to the proposed limit at NH 285257. This second site occurs in an area of knolls and intervening rock basins, located 2.1 km from the downvalley glacial maximum limit and less than 1 km from a clear recessional moraine. It yielded a core of 174 cm depth. A grass-sedge dominated community exists here with *Eriophorum angustifolium* and *Juncus effusus*. *Calluna vulgaris* and *Erica cinerea* thrive on the slopes above the site.

Figures 7.1
Location of pollen site GA I
Figure 7.2
Location of pollen sites GA II and GA III

Plate 7.1
Location of pollen site GA I
7.2.2.3. Glen Affric III

GAIll is situated at NH 327274 at an altitude of 320 m. The basin from which the cores were removed is named Loch na Gobhlaig on the 1:25,000 Ordnance survey map (Plate 7.2). There is, however, little open water remaining in the basin, which is roughly circular in form, and it provides a flat bog surface of approximately 150 m in width. A small stream drains the basin, flowing into the larger Loch Caoireach. The sediment infill attains a maximum depth of 720 cm. Site GAIill is 2.4 km to the north east of the proposed Loch Lomond Readvance limit at Knockfin in the valley of the Abhain Deabahag, where the Glen Affric glacier divides to terminate in two lobes, this being the southernmost one. The area is one of 'knock and lochan topography'. The present vegetation of the basin is dominated by grasses, sedges and rushes, and Sphagnum spp. and Menyanthes are common towards the wetter parts of the basin with rafts occurring in places. Surrounding the basin is abundant Eriophorum augustifolium with Calluna and Empetrum.

7.3. Methodology

7.3.1 Field techniques

At each site traverses were made at regular intervals across the basin and the depth of sediment recorded at points along each traverse. Care was taken to locate the deepest sediment in order that the finest resolution was obtained. It was assumed that the
maximum depth of sediment represented the longest sediment record and that the oldest sediments were therefore sampled. A Russian corer was used at all sites. The 1 m by 5 cm chamber produces a semi-cylindrical sediment sample, 60 cm in length. Duplicate cores were extracted at each site using alternately two holes not more than two metres apart at the surface with a view to maintaining stratigraphic continuity. The procedure for using an instrumental level in order to extract vertical cores and obtain sediment from the correct depths, given by Walker (1974) and Walker and Lowe (1976) was employed. Each core was extruded into polythene covered, plastic semi-circular containers, details of the stratigraphy noted and the core sealed immediately in plastic sheeting. The cores were transported to St. Andrews and stored in a cold room at 4°C.

7.3.2. Laboratory techniques.

7.3.2.1 The Glen Affric, Geusachan and Loch na Gobhlaig cores were removed to a laboratory, still inside their tubes and the surface layers were removed before sampling occurred in order to eliminate any possible contamination. The detailed stratigraphy was noted and the sediment type and major changes in particle size were recorded. The colour of the sediments was also established by means of a Munsell chart.

7.3.2.2. Sampling

The basal sections were sampled as these were of more relevance to the study. Samples from the centre of the cores were taken at intervals of either one or two centimetres. Material was initially removed from the basal sections and was later supplemented, after initial pollen analysis, with additional samples at the respective intervals from the cores, up to levels in which pollen analysis demonstrated that Corylus was clearly dominant. The length of core providing samples was ultimately 25 cm from the shortest core (GAI), 32 cm from the lowest core from GAI and the basal 48 cm from GAIII.

7.3.2.3. Chemical preparation techniques

Absolute pollen counting was employed for each core. The calculation of absolute pollen values requires knowledge of the number of exotic grains that are added to each sample; this is calculated from the use of a tablet containing 10850 grains added to a 0.5 cm³ sample. The samples were prepared for pollen analysis using the following procedures: the sediment samples were measured for 0.5 cm³ by displacement of dilute sodium hydroxide solution in a 5 ml measuring cylinder (Bonny, 1972). A tablet of Lycopodium clavatum spores was then added to each sample in order that the pollen concentration values at each of these levels could be established (Benninghof, 1962; Stockmarr, 1971; see below). Standard procedures for pollen extraction were subsequently employed (c.f. Faegri and Iversen, 1989).

The samples were first treated with HCL (to dissolve the tablet); this was followed by digestion in 10% NaOH to remove humic soil colloids, sieving through 180μm sieves to
remove large particles (and rinsing by deionised water), 40% HF treatment for one hour (Pennington et al., 1972) for removing silica and silicates, 10% HCL, Erdtman's acetolysis for the removal of lignin and cellulose and finally dehydration in 95% absolute ethanol and tertiary butyl alcohol.

7.3.2.4. Mounting
Following chemical preparation the samples were then mounted, unstained in silicone fluid (Andersen, 1969) which has a low refractive index (Berglund et al, 1960) and a high viscosity of 12,500 cst.

7.3.3 Pollen concentration studies
Since radiocarbon dates have not yet been established for the sites described it was impossible to attempt to calculate sedimentation rates. It is hoped however, that dates may be achieved in the future. The absence of dates from all sites precluded the delimitation of chronozones. This restricted the accuracy with which the regional vegetation history could be established.

7.3.4 Pollen deterioration studies
7.3.4.1 Introduction
The state of deterioration of pollen grains may provide information as to syndepositional environmental conditions and may reveal the possible occurrence of redeposition of pollen (Moore et al , 1991). Recent studies have stressed the need for a quantitative assessment of deterioration in pollen stratigraphic analysis. (Walker and Lowe, 1990)

7.3.4.2 Value and aims of pollen deterioration study
The analysis of the state of deterioration of pollen grains is primarily useful, and it may be suggested most necessary, in checking the reliability of the representation of pollen spectra in pollen diagrams (Tipping, 1984, 1987). The proportions of indeterminable pollen are also calculated. Where the percentage is less than 5% it is considered that the palaeoecological interpretation of the spectra will be accurately represented.

The analysis of pollen deterioration may also serve to give a greater understanding of the sedimentological history of a particular pollen site. Deterioration studies may indicate the occurrence of pollen redeposition (Cushing, 1964; Birks, 1970) through high percentages of corroded and degraded grains. Pollen deterioration data may also be employed to detect phases of inwashing of material (eg. Lowe, 1982). Tipping (1984) identified the occurrence of inwashing of material in the Loch Lomond Stadial where lithological information was absent. The nature and amount of pollen deterioration may also reflect
fluctuations in the water table (Lowe, 1982; Delcourt and Delcourt, 1980). Few pollen
deterioration studies have been published. Much remains speculative and further study is
required into the processes of deterioration and how these may be related to
syndepositional processes.

7.3.4.3 Previous work on pollen deterioration
Whereas some palynologists record the nature of deterioration of the unidentified fraction
of the pollen counted, a larger sample is provided for the purposes of analysis of
preservation if all pollen grains are assigned a category of state of preservation or
deterioration. Researchers have derived a number of systems for differentiation of
deterioration types (e.g. Cushing, 1967a; Tolonen, 1980a; Delcourt and Delcourt, 1980).
Cushing (1964) identified only two categories of corroded and degraded and Birks (1973)
added further classes to these two: crumpled and split. The classification proposed by
Delcourt and Delcourt (1980) is however based on a more comprehensive survey of
causes of deterioration.

7.3.4.4 Classification of pollen deterioration
Several classification schemes have been proposed, describing a range of deterioration
types. Combining the previously used hierarchical schemes proposed by Cushing (1964)
for example and the classification of the state of deterioration of pollen grains of Delcourt
and Delcourt (1980), a five fold system was used but modified to differentiate between
crumpled and folded grains: a set of seven classes defined in Table 8.1. All determinable
pollen grains were assigned to a taxon and each grain was assigned to a class relating to
the dominant state of deterioration. Indeterminable grains were unassignable to a taxon
but were categorised in terms of the type of deterioration. Some grains were rendered
unidentifiable due to concealment behind debris.
Grains that exhibited a single fold but whose exine was well preserved were recorded as
'folded', as they would otherwise have had to be classified 'crumpled' which was
considered an inappropriately severe description of the state of preservation. The 7
classes are not mutually exclusive. Grains may exhibit characteristics of more than one
type of deterioration. A hierarchy of deterioration classes was used (i) - (v) and grains
were included in a single class.

<table>
<thead>
<tr>
<th>Table 7.1</th>
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<tbody>
<tr>
<td>(i) corroded</td>
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<tr>
<td>Perforations in the exine, scored etched or pitted surface.</td>
</tr>
<tr>
<td>(ii) degraded</td>
</tr>
<tr>
<td>General thinning of the exine rather than local perforation, Exine elements may be obscured or apparently fused as a structureless mass.</td>
</tr>
</tbody>
</table>
(iii) broken | Part of the grain has been sheared or split to change the whole structural appearance of the grain

(iv) crumpled | Misshapen grains that have clearly been compacted or subjected to other forms of physical stress.

(v) folded | Grain reveals simple fold or bend but exine remains intact.

(vi) obscured | Grain is hidden under detritus which has proved resistant to chemical treatment.

(vii) well-preserved | No or slight visible deterioration

7.3.4.5. Causes of pollen deterioration

(i) Corrosion
A dominant cause of corrosion is that of microbial activity (Havinga, 1964, 1967) in which fungi and bacteria produce local oxidation of the exine, usually under conditions of periodic aeration. This indicates that corroded grains have been in contact with oxygen after deposition on the sediment surface or during a previous period. Corrosion by microbial attack in anaerobic conditions may continue at a slower rate in waterlogged sediments (Clymo, 1965). Havinga (1967) also suggested that different types of microorganisms may result in difference in the intensity of corrosion.

(ii) Degradation
Moore et al. (1991) proposed that thinning of the exine results from exposure of grains to air and associated chemical oxidation and that this process may occur during periodic drying of the surface materials. Alternatively, Cushing (1964, 1967) and Birks (1970) proposed that degradation has a mechanical origin as it is usually associated with minerogenic sediment. This view is supported by the fact that there is little correlation between orders of susceptibility to corrosion and degradation (Tipping, 1984). Burial time has also been invoked to explain the degree of degradation (Faegri, 1971).

(iii) Breakage and splitting
The physical stress required for this to occur in a grain’s depositional history may be achieved by stream transport. It has also been suggested that macrofaunal ingestion may induce crumpling (Lowe, 1982). This form of mechanical deterioration may also be induced during laboratory preparation of examples. Cushing (1964, 1967) suggested that breakage of conifer pollen could be attributed to preparation techniques and this was confirmed by Tipping (1984).

(iv) Crumpling
This can occur in any phase of transport, deposition or storage. High proportions of crumpled grains have however been associated with stream-transported sediment (Birks, 1970; 1973). Post depositional compaction and crushing also occur due to burial under sediment. Faunal ingestion has also been invoked to explain the occurrence of crumpled
grains (Moore et al. 1991). Pragowski (1970) demonstrated that crumpling can be effected by chemical preparation and the viscosity and speed of solidification of the mounting medium, as this affects pressure asserted on grains.

(v) Folding
Minor mechanical damage due to limited physical stress occurring during transport or laboratory treatment.

7.3.4.6. Problems of pollen classification and interpretation
The first difficulty in pollen deterioration relates to assigning a pollen grain to a deterioration class. Various schemes rely on the definition of a hierarchy of deterioration. Tipping (1984) suggests however, that this is inappropriate, as the basis for such a division is unfounded, and his classification is based on the dominant form of deterioration. The heirarchical scheme has remained in general use (Walker and Lowe, 1990; 1991) and is adopted here.

With regard to interpretation of pollen deterioration data, a fundamental problem is that of subjectivity. An analyst attempts to make an objective assessment of each grain but may be influenced in the knowledge that different taxa respond to deterioration to different degrees and are susceptible to different types of differentiation. It is possible that the definition of principally mechanically deteriorated grains changed as each new grain of a taxon was assessed. The definition therefore becomes determined by a comparison with grains previously encountered. It has been demonstrated (Cushing, 1964; Havinga, 1964, 1967; Andersen, 1967) that taxa are prone to differential susceptibility in terms of the nature of deterioration and degree of intensity of deterioration with regard to corrosion and degradation.

There are also problems relating to the interpretation of the data with regard to varying sources of deterioration. Tipping (1984) lists 3 principal periods in which deterioration can occur:

(i) syn-depositional (penecontemporaneous). Deterioration may occur during the time between pollen production and eventual incorporation and burial within a sediment in a basin. This includes processes of transport (Tauber, 1965; Peck, 1973; Bonny, 1976), residence of pollen grains on a soil surface or incorporation within it. Secondary processes of resuspension and redeposition (Davis and Brubaker, 1975) also reexpose grains to contemporaneous deterioration occurring during sediment redeposition at the sediment-water interface, whereas final burial isolates grains from such effects. In the case of grain introduction into peats, pollen may be subjected to aerobic conditions and associated processes of deterioration by a fluctuating water table.

(ii) diagenetic (post depositional). Burial, compaction and compression of the sediment result in pollen deterioration and may promote further deterioration of grains that are already decayed. (Lowe, 1982) Heat and pressure induced deterioration may also occur
(Brooks and Elsik, 1964).

(iii) sampling and laboratory preparation. Research has indicated that pollen deterioration of a mechanical nature is advanced during the chemical preparation of samples (Faegri, 1971), during mounting (Praglowski, 1970) and during the emplacement of cover slips (Cushing, 1964; Tipping, 1984). It has also been suggested that exposure of sediment to fungal or microbial attack during storage may add to mechanical or chemical deterioration.

The relative importance of deterioration in each of these phases is uncertain. Deterioration may occur in one or all of these phases. It is generally considered that most deterioration is a result of penecontemporaneous processes (Faegri, 1971; Lowe and Walker, 1977).

Lowe and Walker (1977) stressed the importance of decay associated with high rates of solifluction and inwash during the Lateglacial. Lowe (1982) emphasises the difficulties of attributing deterioration to particular phases in a discussion of basin edge collapse and redeposition of buried pollen and Tipping (1984) reaffirms such difficulties of classification. It has also been suggested that 'progressive pollen deterioration' may occur with increasingly intense deterioration (Hall, 1981) but there is little evidence to confirm this. The inferred origins of deterioration must be considered in conjunction with causes of deterioration outlined above.

7.3.5. Pollen counting

7.3.5.1. Counting strategy.
Pollen and spore counts were undertaken on a Nikon Research microscope. Routine identification of pollen was made using a medium power (x40) objective enabling a magnification of 600. Critical or difficult identification was undertaken at x1000 under oil immersion. Counting was carried out by means of regularly spaced traverses across the slide. The basic counting sum was 500 determinable land pollen (TLP) but the required total was lowered to 250 and even on occasion 100 where pollen concentration was very low, especially in basal minerogenic sediments. A number of slides, typically three or four, was usually needed to meet the prescribed total although on occasion the total was reached before a slide had been completed. In such cases it was considered necessary to complete the whole slide, in view of the non-random distribution of grains on slides (Brooks and Thomas, 1967). During counting, records were compiled of the state of preservation of pollen grains and spores as described above.

7.3.5.2. Pollen Identification.
The identification of pollen grains and spores was achieved by comparison with a type-slide collection and by reference to Moore and Webb (1978), Erdtman et al, (196; 1963). Faegri and Iversen (1975) and Andrew (1984) were also consulted frequently. Grains were classified according to the key given in Moore and Webb (1978). The nomenclature
adopted was that of Stace (1990). The identification of grains was generally made to the family level in the case of non-arboreal pollen. It was considered advantageous to subdivide certain families into several distinct subgenera such as *Rumex* and *Rosaceae* and in addition to this a *Rumex* undifferentiated or *Rosaceae* undifferentiated class was included. Rarely was it necessary nor possible to identify to the plant species level. The general state of grain preservation and the difficulties of attempting routine identification at a greater magnification prohibited observation of the features necessary for such critical identification as listed in keys such as Faegri and Iversen (1975).

Included in the pollen diagrams are certain elements that merit further comment:

1. **Betula.** Several studies have attempted to differentiate species of this family, particularly between *Betula pubescens* and *Betula nana*. Writers have described how this may be achieved by two sets of measurements.

   In this study *Betula Nana* has been differentiated subjectively. The lack of time available meant that it was not possible to undertake to measure the pore sizes of all *Betula* encountered and whilst it was recognised that the use of HF may have a distorting effect on grain size, measurements of grain size were noted.

2. **Corylus/Myrica.** No attempt was made to distinguish between *Corylus* and *Myrica* gale pollen. Grains are termed *Corylus* but fossil pollen of either and both types may be present.

3. **Salix.** Identification was made to the family level, although various species were noted.

4. **Empetrum.** Differentiation to species level was prohibited by the poor state of preservation by many grains. It would also have required measurement to separate *E. hermaphroditum* from *E. nigrum*.

5. **Myriophyllum.** Much of the pollen observed was *M. alterniflorum* but *M. spicatum* was also identified.

### 7.3.6. Sediment stratigraphy

The depth of sediment from each site was in all cases less than the length of one core (<50 cm). There was, therefore, no problem with regard to matching cores from adjacent cores-holes at a site. The lithostratigraphy at each site is detailed in Table 8.2. (Page 193) The term 'gyttja' is used to indicate a finely-divided limnic deposit derived from the flora and fauna of the eutrophic lochan that occupied the basin. At certain levels the gyttja is purely organic and in other places contains a greater proportion of minerogenic sediment - a clay-gyttja (although silt may be present).

### 7.3.7. Pollen diagram construction

Both relative and absolute pollen diagrams have been constructed for each site. The vertical scale shows the lithostratigraphy of the core and the depth from the surface from which the samples and counted pollen is taken. The taxa identified at each site are represented in the conventional manner along the horizontal axis. The diagrams were
constructed by *Tilia Graph* (Grimm, 1991) and drawn from calculated percentages and concentrations. The pollen diagrams were based on sums of Total Land Pollen. At levels where the percentage calculated was less than 2%, this is recorded by a +.

### 7.3.8 Zonation of pollen diagrams

Zonation of the pollen diagrams was undertaken to facilitate the description of the diagram and the associated interpretation of vegetation sequences. This also enabled correlation of the diagrams to produce Regional Pollen Assemblage Zones and with other biostratigraphic sequences.

Although the zonation of diagrams may be identified on the basis of lithostratigraphy, this technique was not employed as microfossil content does not necessarily coincide with changes in pollen content. The zonation of the pollen diagrams was instead based on the pollen components present to produce a sequence of biostratigraphic units. Initial steps to identify the local pollen assemblage zones were undertaken using the CONISS programme in *Tilia*. This used pollen percentage data to identify units by cluster analysis. The programme used percentage pollen data based on the major taxa present. The programme identifies major groupings of taxa representation and the main obvious breaks between associations may be used to define zone boundaries. Where distinct units were less clear, a more subjective approach was incorporated: ecological, climatic and temporal considerations were introduced to check and identify possible zones and boundaries. Adopting the empirical approach advocated by Cushing (1967b) each diagram was divided into internally homogenous units displaying recognisable pollen characteristics that could be differentiated from adjacent units.

Such zones may represent periods of change or relatively static phases. The boundaries are, however, based on the synchronous behaviour of several taxa although the changes in one or more particular taxa (e.g. *Betula*) may be considered especially significant in terms of reconstructing former vegetation changes and zonation may accordingly be based on fluctuations in such pollen representations. It was possible to subdivide the local pollen assemblage zones into sub zones to highlight vegetation changes at individual sites.

The zones located at each site are local zones but are used to produce regional pollen assemblage zones and to achieve comparison with chronozones and pollen sequences from NW Scotland (Pennington, 1972), the Moray area to the east, the western Highlands (e.g. Walker and Lowe, 1981) and Skye (Walker and Lowe, 1990). Regional pollen assemblage zones were identified using all the diagrams and are discussed in section 7.8.

### 7.4 Results of pollen analyses

#### 7.4.1 Pollen percentage diagrams
The percentage pollen diagrams are presented in Figures 7.3 - 7.5.

7.4.1.1. GAl

The percentage pollen diagram for GAl is given in Figure 7.3. Local pollen assemblage zones were delimited by the *Tilia* programme. These are defined as follows:

<table>
<thead>
<tr>
<th>GAl</th>
<th>Pollen Assemblage Zone</th>
<th>CM</th>
<th>4a</th>
<th>4b</th>
<th>5a</th>
<th>5b</th>
</tr>
</thead>
<tbody>
<tr>
<td>GAl 4</td>
<td><em>Betula - Corylus</em></td>
<td>640</td>
<td>647.5</td>
<td>655</td>
<td>660</td>
<td>663.5</td>
</tr>
<tr>
<td></td>
<td><em>Corylus - Betula</em></td>
<td>647.5</td>
<td>655</td>
<td>660</td>
<td>663.5</td>
<td>669</td>
</tr>
<tr>
<td></td>
<td></td>
<td>640</td>
<td>647.5</td>
<td>655</td>
<td>660</td>
<td>663.5</td>
</tr>
<tr>
<td>GAl 3</td>
<td><em>Betula - Juniperus</em></td>
<td>655</td>
<td>660</td>
<td>663.5</td>
<td>669</td>
<td></td>
</tr>
<tr>
<td></td>
<td><em>Juniperus - Empetrum - Betula</em></td>
<td>663.5</td>
<td>669</td>
<td>672</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GAl 2</td>
<td><em>Empetrum</em></td>
<td>663.5</td>
<td>669</td>
<td>672</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GAl 1</td>
<td><em>Poaceae - Betula - Rumex - Salix</em></td>
<td>669</td>
<td>672</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The four zones described above are shown as GAl 1 - GAl 4 on figure 8.3. GAl 4 has the greatest internal consistency. Greater internal variation is seen in GAl 2 and GAl 3.

The local pollen assemblage zones (LPAZs) are described in turn in terms of composition:

**GAl 1**
This basal zone corresponds to a zone of minerogenic banding. The major contributors to the pollen and spore influx in GAl 1 are Poaceae and *Betula*. Poaceae achieves its maximum value in the upper part of this zone and both *Rumex* and *Salix* peak in this LPAZ. Cyperaceae is also present in significant amounts.

**GAl 2**
This zone is characterised by a sharp increase in *Empetrum*. The double peak towards the upper boundary of the zone is only due to a decrease in one level. Poaceae values are generally maintained at c. 15%. Values for *Betula* decrease to a profile minimum at 667 cm but then begin to increase; *Rumex* also declines in the lower zone. *Huperzia selago* peaks in the lower zone and reaches appreciable counts in the middle part of this zone. *Polypodium* and *Myriophyllum* are also important in GAl 2. This zone also reveals the introduction of *Juniperus*.

**GAl 3**
The lower part of this zone seems to reflect marked fluctuations in the representation of certain pollen taxa, notably *Betula, Empetrum* and *Corylus*. The zone is characterised by high *Juniperus* and *Betula* pollen values. A general decrease in *Empetrum* occurs throughout the zone and *Corylus* begins to assert itself towards the upper levels. Poaceae values also exceed those in GAl 2.

**GAl 4**
Pollen taxa are fairly constant in the lower part of this zone with *Betula* and *Corylus* the dominant contributors to total land pollen. A distinct change occurs between the subzones, identified by a marked increase in *Corylus* pollen and substantial simultaneous
decrease in Empetrum and Juniperus. Filipendula also becomes important throughout the upper zone. Sphagnum also reaches appreciable levels throughout this zone.

**GAI**

Figure 7.4 shows the pollen percentage diagram for this site. Six local pollen assemblage zones are identified:

<table>
<thead>
<tr>
<th>GAI 6</th>
<th>150 - 153.5 cm</th>
<th>Corylus-Filipendula-Athyrium-Filicales</th>
</tr>
</thead>
<tbody>
<tr>
<td>GAI 5</td>
<td>153.5 - 157.5 cm</td>
<td>Betula-Corylus-Filicales</td>
</tr>
<tr>
<td>GAI 4</td>
<td>157.5 - 162 cm</td>
<td>Betula-Poaceae-Filicales</td>
</tr>
<tr>
<td>GAI 3</td>
<td>162 - 170.5 cm</td>
<td>Poaceae-Cyperaceae-Juniperus-Filicales</td>
</tr>
<tr>
<td>GAI 2</td>
<td>170.5 - 172.5 cm</td>
<td>Cyperaceae</td>
</tr>
<tr>
<td>GAI 1</td>
<td>172.5 - 174 cm</td>
<td>Poaceae-Empetrum-Cyperaceae</td>
</tr>
</tbody>
</table>

**GAI 1**

High values of Poaceae (>30%) and maximal values of Empetrum characterise this lowest zone, although both of these taxa decline through GAI 1. A notable feature with regard to changes in pollen curves is a sharp increase in Cyperaceae from 12 to 30%. Values for Rumex acetosa and for Huperzia selago are also maximal here.

**GAI 2**

Cyperaceae dominates this zone, attaining a maximum percentage of over 50%. This is accompanied by diminishing Poaceae and Empetrum values.

**GAI 3**

CONISS divides the zone between 170.5 and 162 into two zones but this is largely on the basis of the curve for Filicales. Otherwise the zone is internally consistent with percentages of Poaceae always exceeding 25% and reaching a maximum of 38%. Cyperaceae is similarly constantly present at typically 15 - 20%. The most notable change in GAI 3 is the expansion of the curve for Juniperus, although at moderate levels (8 - 15%). Asteroideae and Lactucoideae peak in succession in this zone.

**GAI 4**

At the 3-4 boundary, percentages of Betula increase markedly to exceed 40% and those of Poaceae decline although it remains a substantial contributor to the pollen total at these levels. Corylus begins to become important at the onset of this zone and Filipendula values rise noticeably. Peaks in Juniperus and Empetrum occur at 160 cm. A minor peak in Calluna also occurs at the beginning of this zone.

**GAI 5**

A marked increase in Corylus percentages is registered at the opening of GAI 5 together with a sharp increase in Athyrium and Filicales. Betula percentages remain high (25-35%) whilst Poaceae, Cyperaceae and Empetrum decline to minima in this zone. Filipendula continues to increase and Lactucoideae achieves a minor peak.

**GAI 6**

In GAI 6 Corylus percentages peak at over 30% and Filipendula also rises to a
maximum of 29%. Salix also attains a maximum of over 10% and Juniperus is still present. Betula becomes relatively unimportant here. Athyium and Filicales rise to maxima near the top of the zone.

GAIIII
A percentage pollen diagram for GAIIII is shown in Figure 8.5. The zonation programme in Tilia was employed to zone the diagram and six local pollen assemblage zones were identified:

<table>
<thead>
<tr>
<th>Zone</th>
<th>Depth Range</th>
<th>Pollen Assemblage</th>
</tr>
</thead>
<tbody>
<tr>
<td>GAIII 6</td>
<td>660 - 667 cm</td>
<td>Betula - Corylus</td>
</tr>
<tr>
<td>GAIII 5</td>
<td>667 - 677 cm</td>
<td>Juniper - Betula</td>
</tr>
<tr>
<td>GAIII 4</td>
<td>677 - 687 cm</td>
<td>Empetrum - Poaceae - Cyperaceae</td>
</tr>
<tr>
<td>GAIII 3</td>
<td>687 - 699 cm</td>
<td>Empetrum</td>
</tr>
<tr>
<td>GAIII 2</td>
<td>699 - 705 cm</td>
<td>Salix - Rumex - Poaceae</td>
</tr>
<tr>
<td>GAIII 1</td>
<td>705 - 708 cm</td>
<td>Poaceae - Salix</td>
</tr>
</tbody>
</table>

Six clear zones were apparent and, as there was no appreciable differentiation within each zone, it was deemed unnecessary to subdivide them.

GAIII 1
Poaceae attains maximum values in this zone and it constitutes the major contributor of pollen. Salix is important here, increasing through the zone to a maximum in GAIII 2. Cyperaceae values are moderate and Betula is initially important. Maximum values of Rumex undiff. are recorded in GAIII 1, Rumex acetosella peaks for the first time and Rumex acetosa increases here. Lactucoideae and Artemisia peak in the basal sections and fail to achieve such levels higher up the sampled profile. Herbs are otherwise absent from this zone.

GAIII 2
Salix reaches a peak and sustains high values in GAIII 2 whilst Rumex flourishes to a maximum in the upper part of the zone, especially Rumex acetosa but Rumex acetosella also reaches its maximum here. Poaceae steadily declines although remaining a major component of the TLP. Cyperaceae values decrease whilst Betula is fairly constant. There is a rapid increase in Empetrum at the upper boundary of the zone. Artemisia is consistently present throughout this zone. Herbs such as Apiaceae, Caryophyllaceae and Chenopodiaceae also occur at the top of this zone.

GAIII 3
Empetrum is clearly the dominant component of this zone, attaining high levels throughout. Salix declines to a minimum but taxa such as Cyperaceae, Poaceae and Betula show slight increases and Corylus makes an early appearance in GAIII 3. Many more herbs appear in this zone and those such as Filipendula, Caryophyllaceae and Artemisia are continually present here. Calluna is likewise present throughout, increasing towards the upper boundary. Rumex is also present although declining.
GAIII 4
Values of Poaceae remain high, reaching their maximum and dominating this zone together with high, although slightly lower than GAIII 3 values of Empetrum. A third important element is that of Cyperaceae which peaks in this zone. Following an initial expansion and decline in Zones 2 and 3, Juniperus reasserts itself, rising rapidly towards the close of GAIII 4. Corylus continues its presence but decreases in representation at the GAIII 4 - GAIII 5 boundary. Low peaks in Pinus and Lactucoideae also occur in the upper part of the zone. Betula, Salix and Rumex totals remain relatively constant through GAIII 4.

GAIII 5
The major characteristic of this zone is the rapid expansion of Juniperus and its continued dominance into the zone above. This is accompanied by a steady increase in Betula pollen and a decrease in Empetrum, Poaceae and Cyperaceae. Calluna becomes more important and a significant rise in Corylus commences in the upper levels of the zone. With regard to the other taxa, Apiaceae becomes rare, Ranunculaceae become more common and Mentha, Myriophyllum and Dryopteris are better represented.

GAIII 6
As Juniperus gradually declines in GAIII 6, first Betula and then Corylus reach maximum peaks in the middle and upper parts of the zone respectively. Pinus, Salix and Calluna increase from low values at the beginning of the zone.

7.4.2. Lithostratigraphy
The three lithological profiles (Tables 7.1-7.3) appear to postdate the Loch Lomond Readvance. A transition from minerogenic sediment to gyttja is observed in all cases. The fact that there was no Lateglacial organic deposit at GAI and GAI1 is consistent with the reconstruction of Loch Lomond Readvance limits. GAIII was expected to yield such a deposit and its absence may be explained in a number of ways: (i) The corer simply failed to penetrate through the minerogenic layer beneath the sampled organic deposit. The former may have been especially thick due to extensive solifluxion during the Loch Lomond Stadial. (ii) The true Loch Lomond Readvance limit may be outside this site. This is contrary to clear mapped evidence.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Description</th>
<th>Colour/Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>450 - 532</td>
<td>Peat</td>
<td></td>
</tr>
<tr>
<td>532 - 571</td>
<td>Dark brown gyttja</td>
<td>Very dark brown 10YR.2/2</td>
</tr>
<tr>
<td>571 - 589</td>
<td>Brown gyttja</td>
<td>Dark brown 7.5Y.3/2</td>
</tr>
<tr>
<td>589 - 597</td>
<td>Very dark brown gyttja</td>
<td>Black 10YR.2/1</td>
</tr>
<tr>
<td>597 - 605</td>
<td>Grey clay</td>
<td>Dark grey 5Y.4/1</td>
</tr>
<tr>
<td>&lt;605</td>
<td>Grey clay with coarse gravel</td>
<td></td>
</tr>
</tbody>
</table>
Table 7.2

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Description</th>
<th>Colour/Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>114 - 125</td>
<td>Dark brown gyttja with mica</td>
<td>Black 2.5/5Y</td>
</tr>
<tr>
<td>125 - 155.5</td>
<td>Finer dark brown gyttja with mica</td>
<td>Black 2.5/5Y</td>
</tr>
<tr>
<td>135.5 - 140</td>
<td>Dark brown gyttja with mica</td>
<td>Black 10Y.2/1</td>
</tr>
<tr>
<td>140 - 142</td>
<td>Finer dark brown gyttja with mica</td>
<td>Black 2.5/5Y</td>
</tr>
<tr>
<td>142 - 143.5</td>
<td>Dark brown gyttja with mica</td>
<td>Black 10Y.2/1</td>
</tr>
<tr>
<td>143.5 - 152</td>
<td>Finer dark brown gyttja with mica. Birch twig at 151 cm.</td>
<td>Very dark brown 10Y.2/2</td>
</tr>
<tr>
<td>152 - 161</td>
<td>Coarse sand</td>
<td>Dark grey brown 10YR.4/2</td>
</tr>
<tr>
<td>161 - 164</td>
<td>Sand (less coarse)</td>
<td>Dark grey brown 10YR.4/2</td>
</tr>
<tr>
<td>164 - 165.5</td>
<td>Fine sand</td>
<td>Very dark greyish brown 10YR.3/2</td>
</tr>
<tr>
<td>1655 - 174</td>
<td>Silty sand</td>
<td>Dark grey 10YR.3/2</td>
</tr>
<tr>
<td>&lt;174</td>
<td>Grey clay with coarse gravel</td>
<td></td>
</tr>
</tbody>
</table>

A sharp break occurs at 1.52 m. The grey clay with coarse gravel at 1.74 m proved to be impenetrable.
Table 7.3

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Description</th>
<th>Colour/notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>673 - 680</td>
<td>Gyttja</td>
<td>Olive 5Y.4/2</td>
</tr>
<tr>
<td>680 - 683</td>
<td>Gyttja</td>
<td>Dark greyish brown 2.5 - 3/2</td>
</tr>
<tr>
<td>683 - 689.5</td>
<td>Gyttja</td>
<td>Olive grey 5Y.4/2</td>
</tr>
<tr>
<td>689.5 - 693.5</td>
<td>Gyttja</td>
<td>Very dark greyish brown 2.5Y.4/2</td>
</tr>
<tr>
<td>693.5 - 696.5</td>
<td>Gyttja</td>
<td>Dark greyish brown 2.5Y.4/2</td>
</tr>
<tr>
<td>696.5 - 698</td>
<td>Gyttja</td>
<td>Very dark greyish brown 2.5Y.3/2</td>
</tr>
<tr>
<td>698 - 704.5</td>
<td>Gyttja (finer)</td>
<td>Dark greyish brown 2.5Y.4/2</td>
</tr>
<tr>
<td>704.5 - 720</td>
<td>clay</td>
<td>Grey 5Y.51</td>
</tr>
<tr>
<td>&lt;720</td>
<td>Very coarse gravel (3cmx2cm) in fine grey clay</td>
<td></td>
</tr>
</tbody>
</table>

7.4.3. Pollen concentration diagrams

These were produced in order to detect fluctuations in the concentration of pollen of particular taxa for comparison with the pollen percentage diagrams. They allow greater precision in the interpretation of the patterns of vegetation change. The diagrams are labelled 7.6 - 7.8

7.4.4 Pollen deterioration

This section presents the results of analyses of deteriorated pollen for the three sites in Glen Affric. The problems of classifying and interpreting deteriorated pollen have been discussed above. The diagrams 7.9, 7.10 and 7.11 are presented in a format such as that used by Lowe (1982b) which focuses on determinable deteriorated pollen grains. The local pollen assemblage zones defined in 7.4.1 have been transferred onto these diagrams and are labelled at the right hand side of the diagrams. The boundaries are not, therefore, based on fluctuations in deteriorated pollen totals. Four sets of data abstractions are represented in diagrams 7.9, 7.10 and 7.11.

I  Total deteriorated pollen is given as a percentage of TLP.
II  A series of graphs for each class of deterioration shows the varying
representation of each class as a percentage of total deteriorated pollen. (This reveals the possible processes that have taken place at different levels, for example through the relationship of different types of deterioration to lithology or to major changes in local pollen assemblages).

III This section includes only taxa that are prominent contributors to deteriorated pollen totals. The inner curve shows each taxon as a percentage of total determinable deteriorated pollen and the outer curve total deteriorated pollen (determinable and indeterminable). (This highlights the varying proportions of those taxa making the most significant contribution to overall totals).

IV For each of the taxa given in III, frequency curves are based on the sum of that taxon. The deteriorated pollen total for each taxon is shown as a percentage of the total number of deteriorated and non-deteriorated grains and the deteriorated pollen curve is then subdivided to show the varying proportions of each category of deterioration. (A taxon in III may be important simply because influx of that taxon is high. Also, where total deteriorated pollen totals are low, curves in III can appear misleadingly prominent.)

The diagrams are discussed in turn, in terms of trends in pollen deterioration:

GAI (Figure 7.9 I - IV)
Total deteriorated pollen totals are constantly high throughout the profile, ranging from 61% in GAI 1 to 89% in GAI 4. Two distinct peaks in deteriorated pollen occur in GAI, coinciding with the sharp increase of Betula in the pollen percentage diagram and the second in GAI 3a where Juniperus is becoming important and there are peaks in Empetrum and Poaceae pollen totals.

Corroded and crumpled grains make up most of these totals, as revealed in II. Although the curve for corroded grains commences at low levels (< 9%) in GAI 1, it shows a rapid increase to a maximum in GAI 2 with certain fluctuations thereafter and values of up to 50% in 4a. The rapid increase coincides with a sharp increase in Empetrum pollen and the high incidence of corroded grains in GAI 4a is matched by a steady rise in Corylus. The amount of degraded pollen grains is initially fairly high at 20%. The curve fluctuates but does not rise above 30%, only exceeding 25% with peaks in the upper part of GAI 2 and lower part of GAI 3a. The graph for broken/split grains reveals similarly low levels, between 5 and 20%, peaking in 3b and 4a where the Juniperus percentage pollen curve increases, a pollen type that is known to split easily. A peak such as that in GAI 2 in degraded pollen and associated Empetrum peak is also registered in the percentage curve for crumpled grains which dominates the deteriorated grain totals and reaches a maximum of 46% here, otherwise attaining values of 28 - 40%.

Changes recorded in III reflect major fluctuations in the main pollen percentage diagram, i.e. percentages of deteriorated pollen in a particular taxon are high where that taxon is
well represented in the total pollen assemblage.

The analysis of the percentage of deteriorated pollen for each taxon and the type of deterioration for each taxon in IV enables an assessment to be made of the syndepositional processes including redeposition and to explain anomalies in the pollen percentage curves. For *Betula*, the percentage of deteriorated pollen is initially low in GAl 1, rising sharply to a series of peaks and then steadily increasing to GAl 4b. A high degree of crumpling is apparent throughout the profile and grains are initially deteriorated in this way and are broken. In upper GAl 2 there is a sharp increase in corroded grains followed by a distinct peak in deteriorated grains. High levels of corroded grains prevail until GAl 4b when crumpling is again important. Physical deterioration is then important at first until degradation and corrosion from chemical attack are important for the remaining levels, except where this declines in uppermost levels. Corrosion is the dominant form of deterioration of *Corylus* as it rises in the pollen curve and then crumpling is recorded as being important with a peak of both forms of chemical deterioration occurring at GAl 4a. *Salix* is throughout generally well preserved but instances of increased deterioration are particularly significant. Peaks in both corroded and degraded grains occur in level 662 cm which coincides with an increase in the *Empetrum* pollen percentage curve. Physical deterioration is more important at the base (GAl 1) and through GAl 4a and upper GAl 4b. *Juniperus* pollen has a tendency to split easily and the broken/split category is well represented throughout. Grains also reveal signs of degradation. It is changes in the curves for corrosion and crumpling that are considered important. The lower part of Zone GAl 4b has both increased curves for corrosion and crumpling indicating high levels of both chemical and physical damage. The percentage of deteriorated pollen fluctuates considerably, possibly reflecting periodic inwashing (and redeposition) of sediments. Corrosion is the dominant form of deterioration exhibited by grains. A peak in corroded grains occurs in GAl 3 and, more specifically in 3a. This corresponds with the lower and upper of the three *Empetrum* pollen percentage peaks, suggesting that here redeposition has occurred. Sharp peaks in the levels of degraded grains occur in lower GAl-2 and the GAl-4a/4b transition. Between levels with high values of chemical deterioration, crumpling becomes a dominant form of deterioration. Crumpling has affected most grains up to the 3/4 boundary for Poaceae. A sharp peak in crumpled grains occurs at this level. This is followed by a similarly sharp increase in corroded grains. Finally, Cyperaceae has constant high deterioration levels with both physical and chemical deterioration types fluctuating through the profile.

**GAlI** (Figures 7.10 I-IV)

The proportion of total deteriorated to total land pollen is relatively high exceeding 75% throughout the profile, typically between 85 and 95% up to a maximum of 97%. The curve for I fluctuates with no distinct levels in which deterioration is lower. With regard
to the representation of each class as a % of total recognisable pollen, the classes of corroded, degraded and crumpled are all well represented. There are at all levels over 30% corroded grains and the curve is fairly constant. The curve for degraded grains fluctuates considerably with peaks at 158 cm in GAll 4 and the upper part of GAll 3b. The most notable feature in Section II of Figure 8.10 is the sharp increase in corroded grains at 160 cm coinciding with peaks in the Corylus, Juniperus and Empetrum peaks in the pollen percentage diagram. A minor peak in corroded grains occur at 171 cm and another at 166 cm, the former at the maximum of Cyperaceae in the percentage diagram.

The frequency curves for well represented taxa for GAll show markedly fluctuating curves for Betula, Corylus, Salix, Juniperus, Empetrum, Poaceae and Cyperaceae. Considering first Betula, deterioration is greater in 3a and b and in zone 4. The relative contributions of each of the classes of crumpled, corroded and degraded grains are inconsistent. The incidence of broken grains is less changeable with a distinct increase in percentage values in 154 and 158 to over 10%. High levels of degraded grains are recorded. The Corylus curves demonstrate a general increase in the percentage of deteriorated grains up the profile after an initial peak of corroded grains occurs (with some breakage) and a further peak in broken grains at 157 and 158 cm (the 4/5 boundary). These fluctuations are not matched by features in the Corylus percentage curve. The state of deterioration of Salix grains is unusually high. Substantial peaks in corroded grains occur at 155 cm and in degraded grains at 168 and 152 cm. As with the data for Juniper from other sites, the broken/split category is predominant for this taxon although the relative importance of degraded grains must be noted in that they exceed 50% at several levels. The considerable fluctuation may be due to the low pollen counts for Juniperus at certain levels. A more distinct trend is observed in the set of Empetrum curves with levels of deterioration decreasing up the profile and with crumpling being the dominant form of deterioration at first in GAll 1 and 2 interrupted by peaks in degraded pollen at 167 to 159 cm and a phase of sharply increased corrosion above 162 cm. Poaceae reflects a similar trend of increasingly well preserved grains. Crumpling and then degradation are important contributors to the deterioration of Poaceae pollen up to 162 cm and then corrosion becomes more important. Finally, the curve for crumpled grains peaks at the top of the diagram at over 45%. Trends in Cyperaceae deterioration curves show generally low levels of degraded grains but increasing in the uppermost zone. Similarly percentages of crumpled grains are higher in GAll 6 than in previous zones. Corrosion is important in GAll 2 and GAll 3.

GAllI (Figure 7.11. I - IV)
The proportion of well preserved grains (Figure 7.11. I) is generally constant fluctuating more in GAllI 1 and 2. Distinct and notable peaks are observed in all the deteriorated pollen curves (Figure 8.11 II). The level of corroded grains fluctuates markedly in zones GAllI 1 - GAllI 3 with peaks at the 2 - 3 transition and in upper GAllI 3. A peak in
crampled grains (which make up a high proportion of deteriorated grains) at the 4/5 boundary is followed by one of broken grains. A sharp peak in degraded grains precedes this in zone 4, the second and smaller of two peaks the first occurring in lower zone 3.

Figure 7.11.III shows similar trends to those for GAI and GAIL with deterioration curves closely associated with changes in the percentage diagrams for taxa.

The percentage of well-preserved (Figure 7.11.IV) Betula grains fluctuates somewhat but there is a distinct decrease in this category in the upper part of 5 and throughout 6. Levels of corrosion remain low throughout although obvious peaks occur in upper 4 and the 5/6 boundary. Broken pollen grains are relatively insignificant, at no point exceeding 5%. The peak in corroded grains in 4 is matched by a sharp increase to a maximum in the crumpled grain curve. Degraded grains are more common up to zone 5 peaking in the middle of zone 4.

The representation of classes of deterioration of Corylus fluctuate enormously (possibly indicating marked changes in environmental processes operating in the basin). Notable features, however, include a peak in well preserved pollen in lower zone 5. Significant levels of corrosion occur at 5/6 transition and upper zone 4 and are coincident with that for Betula in the first instance. A further peak in corroded grains occurs in upper zone 3. Degraded grains are abundant at a number of levels reaching a maximum peak as Corylus pollen enters the diagram (Figure 8.11.III) but still constitutes a low % of TLP. The effect of crumpling is recorded at high levels throughout, although this element is seen to fluctuate due to sporadic increases in corrosion and degradation possibly indicating changing conditions.

A high proportion of well-preserved Salix grains are maintained throughout the analysed profile with most values > 45%. Corrosion of grains is only apparent in very low percentages and breakage is similarly unimportant, although distinct peaks occur around the 4/5 boundary. Crumpling is of varying significance, fluctuating considerably from 0% to >47% in upper zone 4. The percentage of degraded grains peaks immediately after this at the 4/5 boundary.

Well preserved Empetrum grains are well represented in 1 and throughout zones 5 and 6. Major features in the Empetrum curves consist of a maximal peak in corroded grains in upper zone 3, extensive crumpling in upper 4 and lower 3 reflected in sharp peaks. Degradation is moderate and significant in zones 1 - 3 and in zone 6.

Corrosion and crumpling are the dominant states of deterioration of Calluna throughout, the only notable features being a sharp increase in broken grains at 680 (upper zone 4).
and a high degree of crumpling in the middle of zone 3.

*Juniperus* is characterised again by splitting of typically >50% of grains. The decrease in this curve at the top of zone 3 coincides with the onset of *Juniperus* in the percentage diagram and also with an increase in the crumpled and degraded component. Relatively high degradation is recorded in 5 and 6.

Significant changes in the Poaceae curves are few. The crumpled category typically contributes >50% with minor peaks in upper 5 and middle 4 in corroded pollen. In both broken and degraded categories peaks are registered in the middle of zone 3. Cyperaceae curves fluctuate significantly. A major feature is a high % of crumpled grains up to the middle of Zone 5 and thereafter corrosion becomes dominant at the top of the profile.

### 7.5. Interpretation of the pollen deterioration data

Two major applications of the pollen deterioration data are addressed. First, as a means of assessing the reliability of pollen spectra and diagrams and as a tool for analysing anomalies in pollen percentage diagrams prior to making a biostratigraphic interpretation. Second, to elucidate the sedimentological history of each pollen site. A third use of such data is to add to the understanding of deterioration of pollen with regard to intensity of deterioration, susceptibility of taxa, the relationship between deterioration types and intensities and environmental processes and the use of pollen deterioration data in pollen studies.

#### 7.5.1 Problems in interpretation

Levels of deterioration are uncommonly high (>70%) in comparison with data from other sites where deterioration of early Flandrian pollen is typically 20 - 30% of total land pollen, falling from 50 - 60% in the Loch Lomond Stadial. The reason for such substantial differences may be site related or may be a result of methodological variations. In considering possible discrepancies in counting and classification it has to be pointed out that classifications of deterioration is very subjective in terms of the degree of deterioration. Whereas one analyst may class a slightly damaged grain as 'well preserved', another may assign it to a particular deterioration category. In this study it was discovered that many grains, although free from exine damage and thus insufficiently physically damaged to be described as crumpled, were folded. Using the previously determined classes many grains were being assigned to the 'crumpled' class when they merely displayed a single and often slight fold. Following the discovery of this problem of classification, rendering a higher proportion of grains 'deteriorated' where
they might otherwise have been 'well preserved', a note was taken of the incidence of folded grains in the 'crumpled class'. It is considered that the inclusion of such folded grains in this class may be too rigorous and they could equally be described as 'well preserved' thus increasing the percentage of well-preserved grains. It is then possible that the 'well preserved' class may be underrepresented whilst the 'crumpled' class over represented. As this trend is observable at all sites it is probable that this methodological problem is the cause of high deterioration values. It is unlikely that local conditions at all of the sites were sufficiently severe to cause the high levels of recorded deterioration. The relatively low percentages of indeterminable grains confirm this. Alternatively, the laboratory preparation techniques may account for a significant proportion of deteriorated grains due to the failure to use an autostirrer or to the severity of the chemicals on grains.

A second problem is that in order to make a valid comparison of deterioration intensity, the definitions of deterioration types would have to be altered for each taxon. This is a rather subjective exercise and an attempt was made to be rigorous with regard to any signs of deterioration. Taxa are also susceptible to different types of deterioration in varying degrees of intensity. On account of this, taxa such as juniper, which splits easily exacerbate the problem described above of over-representation in the deteriorated totals and the 'broken' class where Juniperus occurs.

Concerning the order of susceptibility, there is a need to determine this initially to ascertain which taxa are indicators of processes in order to attempt any interpretation of former basin environments.

The fact that a hierarchical scheme was used in this study means that interpretations must allow for this. This was considered useful as high levels of corrosion and degradation mean high degrees of deterioration. It is also more likely that corrosion and degradation reflect basin processes whereas it has been demonstrated that crumpling and breakage may be related to preparation of the samples.

It is difficult to determine the causes of deterioration and thus to gain a clear understanding of basin processes. Accounting for the deterioration is problematic due to there being many possible causes for deterioration.

7.5.2. Interpretation of the data for the Glen Affric sites

An interpretation of the trends in deterioration of taxa, total deterioration, and deterioration ratios is attempted in order to determine the reliability of the percentage diagrams and to elucidate the stratigraphic history of the sites, especially with regard to water tables and slope processes. The relationship between sediment stratigraphy and
deterioration patterns is also discussed.

GAI
High deterioration totals are probably due to high incidence of crumpled grains as described above. Peaks in corroded grains at 667 cm and at 664 cm in both deteriorated and crumpled grains may be a result of secondary deposition (Cushing, 1964) and the first and third Empetrum peaks in the pollen percentage diagram coincide with these high levels of corrosion and may not truly reflect local vegetation. A peak in Corylus correlates with a corroded peak and may be a false peak. Evidence for ameliorating conditions in sequence of corroded, then degraded and then crumpled grains is recorded in GAI.

GAIL
Corrosion is more important at GAIL. This is to be expected as the site is shallower and more susceptible to fluctuations in the water table. The marked fluctuations in the curves for both degraded and corroded grains attest to this. It is unlikely that secondary deposition is important in this case.

GAILI
Evidence for ameliorating conditions in sequence of corroded, then degraded and then crumpled grains is recorded in GAILI, with a decrease in the importance of corroded and degraded grains and an increase above zone 3 in crumpled grains.

Comparing trends between sites, the lowest levels are characterised by relatively low levels of corroded and degraded grains. Crumpling is dominant in uppermost levels. The general decline in severity of deterioration up the profile (except for lowest levels) may indicate improving or more stable conditions with less inwashing of sediments. Progressive deterioration was considered not to be operating here: there is no general trend of increasing deterioration totals with depth.

Trends in deterioration coincide with biostratigraphic boundaries to some extent as certain taxa display particular tendencies to contribute to a particular deterioration class. Representation of taxa to deterioration levels also reflect changes in the dominance of certain taxa.

It has also been suggested that basin size determines the degree of deterioration with cores from larger basins registering a greater intensity of deterioration. This appears not to apply for the sites in question: local water tables may account for the major differences.

7.5.3. Trends in pollen deterioration

Pollen deterioration totals are consistently high, reflecting methodological problems outlined above. Trends in deterioration types up the profiles reflect changes in the
dominance of certain taxa, notably Poaceae, Empetrum, Juniperus and Cyperaceae. With regard to the incidence of particular types of deterioration there are high levels of corrosion registered in GAI 2 and GAIII 3, in Empetrum and Empetrum - Poaceae - Huperzia zones respectively. Peaks in Empetrum percentages correspond with peaks in corroded grains from GAI and GAIII and may therefore be associated with secondary deposition. An upper peak in percentages for corrosion in GAIII (Figure 8.8) is associated with a short - lived increase in Corylus prior to the Corylus rise and may also reflect the occurrence of redeposition. Increases in percentages of broken grains are in almost all case associated with Juniperus, which is known to split readily. Such increases occur at 660 cm in GAI, 164 cm in GAII and is most apparent at 674 cm in GAIII where there is a distinct peak in Juniperus. Peaks not related to the Juniperus curve occur at 160 cm in GAIII (which may be more a result of increases in Empetrum and Corylus), 154 and 156 cm in GAII which may be related to increases in Corylus. High percentages in crumpled grains are usually associated with dominance of Poaceae, Cyperaceae and Corylus and in the percentage profiles. In GAI at 678 - 680 cm high levels of crumpling correlate with an Empetrum peak and Cyperaceae peak. In GAI at 150 cm it equates with high Corylus values and at 172 - 3 cm with high Cyperaceae. This also occurs in the profiles for GAI, at 648 cm and is related to Betula and Poaceae peaks. GAII has a greater overall proportion of degraded grains. Peaks at GAIII 158 cm and GAI 664 cm are related to increases in Corylus. Elsewhere increases in Poaceae at 161 - 165 cm GAI, in Juniperus in GAIII at 682 cm and in Empetrum and Cyperaceae at GAIII 694 - 96 cm may be related to peaks in degraded grains.

Empetrum is considered an important environmental indicator: palaeoecologically speaking, and the Empetrum data are especially significant in terms of the proportion of deteriorated to well preserved pollen grains. Many of the grains in GAI II and III exhibit exine damage with levels of high percentages of corrosion and degradation reflecting abrasion by mineral material during inwash from catchment soils or basin margin situations. This suggests that the increase in Empetrum that is apparent in % pollen diagrams is a result of secondary rather than primary deposition.

In comparisons with other pollen deterioration studies, deterioration totals are greater than reported elsewhere for the reasons described above. Conversely, the number of indeterminable grains is low.

7.5.4. Discussion

The primary cause of deterioration is considered to be an unavoidable product of the preparation techniques. The high level of deterioration may alternatively be indicative of an environment being disturbed by deglaciation and paraglacial slope processes. Trends
in several diagrams however reveal evidence for secondary deposition, particularly related to the *Empetrum* dominated zones. Fluctuations in local water tables have also been detected for GAI. Causes and sources of deterioration are otherwise difficult to pinpoint. The percentage pollen diagrams are considered to be accurate apart from the local anomalies mentioned: levels of indeterminable pollen are low and the pollen sequences consistent.

The evidence for a revertence noted in the percentage diagrams is not reflected in the deterioration diagrams. Contemporaneous environmental conditions in terms of increased solifluction with the occurrence of taxa indicative of disturbed ground do not appear to correlate with deterioration intensity or type.

### 7.6 Biostratigraphical Interpretation

Interpretation of the local pollen assemblage zones, defined in 8.4 are required to establish the nature of vegetation development at each site and to devise a series of regional pollen assemblage zones for Glen Affric. Local vegetation conditions and prevailing environmental conditions may be inferred using the descriptions detailed in the previous section, relating to the composition of and trends in component taxa. An accurate interpretation can only be achieved if undertaken in the knowledge of possible areas of error or difficulties involving the relationship between pollen counts and the vegetation it is assumed to represent. The close sampling of the cores shows the nature of plant colonisation and vegetational succession following the wastage of Loch Lomond Readvance glaciers at each site.

#### 7.6.1. Problems of Interpretation.

It is possible to allow for certain problems and steps may be taken to minimise others. Additionally, an interpretation must acknowledge the presence of unavoidable error. The first problem over which there is little control at the interpretation stage is that of representative sampling. The TLP for a site is the final sample in a succession of sampling procedures, the first of which involves choosing a site. Although it is possible to calculate confidence limits for example at the 90% level for each result (Faegri and Ottestad, 1948; Bonny, 1971) and to express each value as a range on the pollen diagram, it was considered of limited use with regard to this study. It is acknowledged, however, that the biostratigraphic interpretation is subject to such sampling error.

Another possible problem is that of mixing of sediments by frost-heaving in frost-susceptible sediments. Such sediments are uncommon in the cores collected in this study but some basal sections may have been prone to frost-heave processes. It is
probable that the presence of severely physically deteriorated grains may be a result of such processes and this is addressed in the interpretation.

Stratigraphic disturbance also occurs due to factors such as faunal activity or adhesion of sediment to ice floes which become inverted (Nicholas, 1967). A more likely source of mixing, however, is the resuspension, redistribution and redeposition of sediment by lake currents (Davies et al. 1973) or by streamflow (Tipping, 1987). This problem is addressed by attempting an interpretation with reference to deterioration data: the anomalous presence of corroded or degraded grains may attest to the occurrence of these processes (Tipping, 1987). This highlights the importance of deterioration studies to the interpretation of vegetation sequences.

The question of pollen deterioration is not only important with regard to redistribution in the profile. The differential susceptibility of types of pollen grain to chemical, biological and physical deterioration may be a significant source of error in the pollen counts observed. It is possible that certain types of grain are underrepresented due to a) complete breakdown or b) rendered indeterminable due to features becoming amorphous or the grain having been deformed beyond recognition.

A more fundamental difficulty arises in relation to the degree to which pollen influx at each level (represented more or less accurately by sampling) reflects the composition of the existing vegetation at and around the site at the time of deposition. The variability of input of pollen into the site decreases with distance from source to basin. The pollen record includes a regional component with grains from up to thousands of kilometres distant as well as the local and extra-local influx.

Patterns of dispersal must also be considered in terms of observed pollen records. The influence of air currents passing over and through forests is important. Regional components include grains from anemophilous pollen such as Pinus. The occasional grains of Ulmus and Quercus are also attributed to long distance transport. The prevailing wind direction must clearly be borne in mind. Insect pollinated plants such as those in the Asteroideae tend to produce large ornate grains and are more likely to be of local origin.

A related problem concerns the amount of pollen produced by certain plants. Plants that produce anemophilous pollen tend to yield large quantities of pollen causing over representation in a diagram. Many studies of modern pollen have attempted to determine the relationship between vegetational composition and local pollen influx (e.g. Birks, 1973). Some elements of the vegetation may not be represented at all due to very low pollen production or ineffective dispersal. It is also possible that the analyst may fail to recognise certain grains. Deterioration may also reduce the accuracy of identification
and reduce the number of species identified.

The relative contributions of local, extra local and regional pollen may also change through time. Variables such as the prevailing wind and vegetational change, for example the immigration of trees, can have a profound effect on the composition of the pollen rain, especially with regard to the regional contributions.

A final important consideration in reconstructing past vegetation concerns the making of inferences regarding the composition of vegetational assemblages on the basis of modern analogues. Whilst modern pollen rain studies may ultimately promote an understanding of pollen floristic assemblages, the change in time, and usually space that is involved in Quaternary sites is problematic.

The diagrams are assessed separately and then compared in the following sections. The focus is on the pattern of immigration and the discussion attempts to elucidate differences between sites in terms of the relative importance of local differences, the regional pattern and influence of glaciation on the sites.

7.6.2 Vegetation development at GAI

Initial postglacial assemblages are dominated by herbaceous taxa, notably Poaceae, Cyperaceae and Rumex. Betula appears to be a moderately important component of the vegetation, probably in the form of the shrub B. nana. Betula includes a range of species that varies in stature, including dwarf and tree species. Further evidence for stands of shrub vegetation is indicated by the continuous and stable presence of willow and sporadic presence of juniper. The Betula pollen may be of long distance origin and of B. pubescens form but this is less likely as the taxon is consistently and significantly present. A long distance origin is however invoked for the occurrence of pine in the pollen spectra at this stage. Included in the pioneer communities are Salix and Rumex acetosa. Pioneer species of sedge are also dominant in these basal sections. The flourishing of Rumex is associated with a discontinuous vegetation cover and immature soils. A relatively rapid transition occurring in the vegetation is indicated. This is discussed in section 7.7. Caryophyllaceae, and Plantago maritima occur here and are also indicative of unstable conditions, whilst Filipendula, and Equisetum attest to fairly wet conditions.

An Empetrum heath vegetation establishes itself in zone GAI 2. High values are recorded in the lower part of zone 2 and a second peak occurs in its upper portions. This first anomalous Empetrum peak may reflect reworking of sediments as seen in the high values of corroded grains. Poaceae initially remains important in the local vegetation
community but declines slightly and there is a sharp increase in *Huperzia selago*, which remains throughout the zone. *Huperzia selago* is a heliophyte capable of growing on a wide range of substrates in mountain areas including solifluction slopes (Dahl, 1957). Pennington *et al.* (1972) also suggested that pollen counts for this type tend to increase with the extent of bare rock surface exposed. *Juniperus* begins to assert itself in the vegetation composition in GAI 2 and it appears to indicate that *Empetrum* and juniper coexisted in this zone. They may have grown at different altitudes or on varying soils or the *Empetrum* peaks may reflect secondary deposition from earlier vegetation perhaps from seasonal meltwater floods from mountain snowfields. It is considered, however, that large quantities of *Empetrum* in the pollen profile probably reflect abundant local *Empetrum* (Birks, 1972; Robinson, 1977).

It is inferred that tree birches were not present in the lower zones until the upper part of zone 2 at which point values increased fourfold and it is suggested that they arrived by migration and rapidly expanded to form woodland which replaced *Empetrum* heath and led to the subsequent decline of juniper. *Empetrum* acts as a pioneer dwarf shrub in open mineral soils, spreading rapidly to form a thick mat so that other woody plants such as *Calluna* are able to enter the succession (Gimingham, 1972). A sharp increase in *Betula* and a short period in which *Corylus* flourished and then declined occurs at the beginning of zone GAI 3. This is accompanied by a sharp decrease in *Empetrum* at one level. The fact that *Corylus* rises is important and may reflect a climatic amelioration or localised and short-lived immigration. Both *Juniperus* and *Betula* peak in this zone indicating the dominance of the birch and juniper, whilst *Empetrum* declines. *Corylus* is continually present although at low levels.

The rise of *Corylus* occurs relatively soon (in terms of its occurrence in the pollen profile) after the rise of *Betula*, having been present at low levels in the fern-rich birch woods previous to this. *Empetrum* and *Juniperus* fall to very low levels, whilst grasses and sedges fall to minima for the profiles as the birch-hazel woodland rose to dominance. *Sphagnum* now becomes more abundant. Maximum values of 60% for *Corylus* are attained in the uppermost levels and hazel is becoming increasingly dominant in the woodland.

### 7.6.3 Vegetation development at GAI II

Various pioneer species recorded at both GAI and GAIlll are missing here. The postglacial vegetation sequence appears to start with an *Empetrum*-Poaceae phase, immediately followed by one in which Cyperaceae is dominant. *Rumex* and *Huperzia selago* provide minor peaks in the initial stage, both of which are hardy pioneer plants indicative of unstable ground. The *Betula* present at low levels may be either *nana* or *pubescens* and is considered to be a dominant part of the vegetation at this stage. Poaceae
is important throughout the profile. Other herbs such as Asteroideae and Lactucoideae, Caryophyllaceae, Chenopodiaceae and Plantago spp. occur here in GAII 1 and are also indicative of unstable conditions.

The marked expansion in Cyperaceae, a component of the local vegetation, is accompanied by an increase in Pinus, probably of long distance origin. Its increased presence may not be significant in the vegetational succession. Lactucoideae becomes important at this level and is a significant constituent of the local assemblage together with Asteroideae. Ranunculaceae, especially Thalictrum, appear at this point and continue through the period represented by pollen samples. Myriophyllum begins to assert itself indicating the onset of wetter basin conditions.

At the decline of Cyperaceae, a series of marked changes occur with the introduction of juniper at moderate but consistent levels. This is probably of regional importance and is accompanied by increases in Poaceae and Filicales. Plantago is discontinuously present in this zone and the increasing presence of Rumex and decreasing Juniperus is not replicated elsewhere.

Within the LPAZ GAll 3, Lactucoideae and Asteroideae attain maxima in turn. As Pinus increases and a first Juniperus peak is reached Filipendula starts to assert itself. Cirsium, Sedum and Epilobium are present for the first time and again may suggest moist conditions.

The arrival of Betula appears to be slow, rising gradually in GAll-2 whilst willow and juniper continue to form a shrub vegetation but Poaceae and Cyperaceae are still dominant. Pinus also appears to be closer. The herbs present now include Apiaceae, Thalictrum, Rubiaceae, Potentilla, Rumex acetosa together with Myriophyllum reveal an assemblage indicative of wet conditions. Very high levels of Polypodiaceae and a continuous Lycopodium presence also occur.

A sharp increase in Betula and the appearance of Corylus with increased Pinus suggests the imminent arrival of more thermophilous tree species. Juniperus continues as part of the vegetation, however, suggesting that it is not replaced by birch and hazel immediately. Alongside the first patches of woodland it is probable that tall wet herb meadow thrived as suggested by the abundance of Filipendula and the presence of Urtica and Valeriana and many other herbaceous taxa such as Ranunculaceae and Geum.

Finally in LPAZ GAll-5, Betula continues to assert itself and the decline of Juniperus coincides with the expansion of Corylus in the area. A slight decline in the birch-hazel woodland is indicated with Poaceae and Cyperaceae remaining stable in the vegetation. The arrival of Corylus in the area appears to occur in close succession to that of Betula.
which it slowly replaces. *Salix* and *Juniperus* remain a significant component of the vegetation. *Calluna* begins to increase in presence and *Empetrum* decreases. Lactucoideae also increases in importance. The abundance of *Filipendula* and associated taxa attest to the continued and increasing importance of a local tall herb meadow community at GAILI. The presence of *Filipendula* is probably related to a local stand; its abundance due to the fact that it usually occurs in dense stands.

### 7.6.4 Vegetation development at GAILI

Given the geomorphological interpretation which placed the site outside the maximum limit of the Loch Lomond Readvance, the pollen profile at GAILI did not yield the expected Interstadial - Stadial - Postglacial sequence of vegetation.

The interpretation at the lowest level of the core is least reliable being based on lower pollen counts. The pattern of vegetation development in GAILI 1, however, reveals the re-establishment of vegetation following the Stadial. It begins with the hardiest pioneer plant types that could have survived in close proximity to the ice. The assemblage at this time includes *Huperzia selago*, *Rumex*, *Artemisia*, Poaceae, Cyperaceae and Caryophyllaceae many of which are species typically found in the Lateglacial of northern Scotland (Pennington, 1978). Heliophytic herbs are represented: *Rumex* is a recognised coloniser of unstable, disturbed ground. It is very capable of rapid re-establishment of its root system following disruption by frost heave (Churchill and Hanson, 1958). *Artemisia* is also associated with substrate disturbance (Matthews, 1970) and is well represented. The plant prefers drier continental conditions. Annuals are less commonly represented as the severity of the environment demands plants with lower flowering response times, which is greatest in perennials.

*Betula* is present but at low and fluctuating levels. It is considered not to be a dominant member of the vegetation as established birch woodland is believed not to have arrived until GAILI 5. Like *Betula*, *Salix* includes a range of species from dwarf to tree forms. The most likely species to be represented here are those which prefer soils with impeded drainage such as *S. parpur*, *S. caprea*, *S. cinerea*, *S. nigricans* and *S. phyllicifolia*. *S. repens* and *S. lapponum* which are shorter forms and *S. herbacea* and *S. reticulata* which are dwarf shrubs may also be represented. Dwarf *Salix* can also survive on disturbed soil. *Huperzia selago* occurs at this stage in the vegetation development. Of the aquatic plants only *Myriophyllum* is important in the basal levels.

Following the initial colonisation, *Rumex* and *Salix* flourished and GAILI 2 is characterised by *Rumex*, *Salix* and *Poaceae*. The occurrence of such a peak in *Rumex* is reported elsewhere and is an important taxa for establishing the nature of glacial retreat (cf. section 7.7.) *Rumex* and *Salix* remain important through the zone, declining
eventually to be replaced by immigrating and more competitive plants. *Salix* reaches a maximum and declines almost totally before the increased presence of *Juniperus*.

Pollen of the dwarf shrub *Empetrum* dominates the next phase of vegetation. It is considered to be of *E. nigrum* type and is abundant in northern Scotland in the Lateglacial and early Flandrian. It is heliophilous and chionophobous but has a wide ecological range on stable, non-waterlogged soils (Dahl, 1957; Hafsten, 1963). The production of litter under the preferred oceanic conditions leads to the formation of an acidic soil and climatic and edaphic conditions were thus favourable for the development of *Empetrum*. High frequencies of the species in the early Postglacial pollen record have been recorded elsewhere and attributed to the oceanic climatic conditions that characterise NW Europe with low summer temperatures, low sunshine hours and a small annual temperatures range, together with a high relative humidity. *Empetrum* often favours sheltered situations on for example, N or NE facing slopes or in the shelter of taller plants. A marked fluctuation in *Empetrum* towards the lower levels of GAIII 3 occurs, suddenly and substantially decreasing with a corresponding increase in *Poaceae*, *Artemisia* and *Rumex*. this may represent a short - lived climatic revertence. It is possible, however, that the incidence of crumpled *Empetrum* tetrads corroborates evidence for increased inwash at this time.

Certain new taxa occur in GAIII 2: *Filipendula*, which occurs in poorly drained habitats such as marshes, tall herb meadows and river banks becomes significant. Iversen (1954) noted that a rise in *Filipendula* such as this is caused by a rise in temperature. *Salix* and *Juniperus* may have formed localised thickets in a landscape of open grass sedge heath, rich in herbs including *Huperzia*.

In GAIII 4 *Empetrum* declines slightly together with *Poaceae* and *Cyperaceae* asserts itself. *Juniperus* is continuously present in levels 3 and 4 but flourishes at the decline of *Empetrum*, at the lower boundary of GAIII 5. An increase such as this in *Juniperus* is generally regarded as an indication of climatic amelioration causing increased flowering and pollen production from tall dense juniper scrub. The occurrence of juniper pollen in a diagram is believed to represent the local presence of the plant as pollen dispersal is ineffective.

Following this, *Betula* begins to become more important and *Juniperus* to gradually decrease. It was considered that prior to this level the smaller dwarf forms of *Betula* were more important with some tree pollen of long distance origin whereas the increase in GAIII 5 reflects mostly regional tree pollen. As *Betula* spread into new territory, seed dispersal occurs ahead of the established woodland. The characteristic Lateglacial taxa such as *Huperzia selago* and *Rumex acetosa/acetosella* persist until the expansion of birches and formation of woodland. *Typha* occurs here and provides some indication of
temperatures of this time: it is thermophilous, with July temperatures at 14°C. The presence of Ericales may be related to the immigration of birch. Some of the taxa may be *Vaccinium* and may have formed an understorey in birch copses (Burnett, 1964; Robinson, 1977). *Corylus* appears to rise very quickly following the *Betula* maximum. The *Ulmus* recorded in GA III 6 is almost certainly not locally present as it is only recorded in low levels.

Several general comments regarding the profile as a whole may be made. *Myriophyllum* is continuously present throughout the profile in low quantities although intermittent in value. Most of this is *Myriophyllum alterniflorum*, an obligate aquatic indicative of the presence of open oligotrophic water. *Potamogeton* is not well represented, encountered only in one lower level.

### 7.6.5 Comparison of vegetation development between sites

The early Flandrian sequence is analysed in terms of consistency of pollen types and synchronicity of environmental determinations. Comparison of the three pollen sites is hindered by the unequal lengths of core up to the *Corylus* rise. The pollen record is less precise in GAI (although it was sampled more closely). The evidence for a vegetational succession from a landscape initially dominated by open habitat taxa to dwarf shrub to one dominated by birch and hazel woodland is contained in about 32 cm of gyttja at GAI, 25 cm at GAIL and 48 cm at GAILI. The most complete pollen sequence is from GAILI.

A *Rumex* peak is identified in GAI 1 and GAILI 1 and 2. There is a very much less pronounced *Rumex* maximum at GAIL and the taxon is of limited importance here. Values for *Rumex* pollen rise to a maximum in both GAI and GAILI.

The minor peak in *Empetrum* values in GAILI 1 is followed by a revertance registered in GAILI 2 prior to its marked expansion. This pattern is not reflected in the lower zones for either GA I or GAILI. In other parts of the diagram, anomalous fluctuation in *Empetrum* are attributed to secondary deposition. Values for chemically corroded or degraded pollen grains are not substantially higher in Zone GAILI 1 and 2 and secondary deposition is probably not occurring here. Other evidence for a revertance episode is apparent: the decrease in *Empetrum* may be associated with an increase in open habitat taxa as well as those indicative of disturbed soils. The presence of Caryophyllaceae in GAILI 2 attests to this together with increases in *Rumex* and *Huperzia Selago* and a maximum for *Artemisia* also suggest a climatic revertance. A climatic revertance is not universally present in the Glen Affric sites although signs of one revertence are present in the lower GAILI zones but its significance may be difficult to prove given the low pollen sums.
A clear *Empetrum* maximum is observed in all the Glen Affric sites. Whereas there is a distinct rise to a substantial maximum in GAI 2 and GAIII 3, *Empetrum* values for GAII are initially maximal, at lower values and at once decreasing. The dominance of *Empetrum* in terms of percentage representation in the diagram and its relative duration in the profiles varies: in GAI the increase to and decrease from the maximum is fairly rapid, and in GAII and III a gradual decrease may be noted, following a swift increase in the case of the latter.

There is some variation in the relative importance of juniper at sites in Scotland in terms of its abundance and timing. (Tipping, 1986). A clear but low peak occurs at GAI 3, several unpronounced peaks are observed in GAIII and in GAIII 5 there is both a pronounced and significant peak. In comparison with other sites *Juniperus* is poorly represented. There is a distinct under representation of *Juniperus* at GAI and GAII which may be due to high deterioration and complete breakdown of pollen grains or to a lack of shelter in the area.

At none of the sites was a clear *Betula* zone defined. In GA I and GAIII, *Betula - Juniperus* and *Juniperus - Betula* zones occurred respectively whilst in GAIII a *Betula - Poaceae* zone preceded the *Betula - Corylus* zone common to all of the Glen Affric sites. Marked increases representing rapid immigration of birch into the juniper scrub occurred markedly in GAI but less so at the other two sites. Respective rises in birch and hazel occur at all sites. The maximum for *Corylus* at GAIII occurs rapidly after that of *Betula*; in GAII there is a longer period of birch fluctuations before the *Corylus* rise and at GA I the *Betula* and *Corylus* maxima are separated by a long period of interplay between the types.

The pollen profiles appear to precede the arrival of *Pinus, Alnus, Ulmus* and *Quercus*. Their representation in the diagrams is variable. High *Pinus* values at the onset of the *Empetrum* rise in GAI 1 and into GAI 2 may correlate with a similar peak in GAI 1 - GAI 2. *Pinus* is otherwise present but in very low quantities through the profiles until the early stages of the rise of *Corylus* in GAI and GAIII, and in GAIII only exceeds 5% around the *Corylus* maximum. Its presence here in GAIII might indicate the arrival of pine stands in the area although it is considered by some writers (e.g. Bennett, 1984) that 15% pollen is the minimum from which local pine can be inferred. Some of the anomalous tree pollen may be due to reworking (Cundill and Whittington, 1986).

Site GAI is distinctly wetter as indicated chiefly by the presence of *Filipendula*. Differences also occur in the pattern of immigration. This is considered in section 8.7. A cool initial period may be indicated in GAI and GAIII. Changes in aquatic flora such as *Myriophyllum* reflect decreasing water levels.
The major influences on the vegetation patterns in Glen Affric may be discerned as location, in terms of immigration and climatic influence, proximity of glaciers and deglaciation pattern and a range of local, site specific factors such as aspect, altitude and shelter. The local, extra-local and regional pollen components can be discerned to some extent. The vegetation in the lower zones has more of a local influence with some regional components identifiable. Following this, the profiles tend to reflect the regional vegetation to a greater extent with local vegetative characteristics also identifiable.

Previous work has shown that it is desirable to integrate sites into a series of pollen assemblage zones (RPAZs) which have regional applicability. Insofar as the pollen record at each site reflects the sequence of local plant colonisation, and on the assumption that the expansion of vegetation at the beginning of the Flandrian was rapid and orderly in terms of sequential immigration of herbaceous species through to arboreal taxa, then the three LP AZs may be integrated into a scheme of RPAZs for the area. A series of regional pollen assemblage zones are therefore derived from the three local pollen assemblage zones. The local pollen assemblage zones for the three sites are given in Figure 7.12 together with the derived Regional Pollen Assemblage Zones for Glen Affric GAR 1 - 5.

7.6.6 Summary of the vegetation development in Glen Affric.

An initial period of pioneer vegetation in Glen Affric included open habitat taxa such as Poaceae, Cyperaceae, Rumex, Caryophyllaceae, Lactucoideae, Asteroideae, Artemisia and clubmosses such as Huperzia selago. Of woody plants only Salix was significant. The plant succession continued with a landscape of dwarf shrub heath and open grassland, the dominant heathland elements being juniper and Empetrum. The continued presence of heath grasses and sedges indicates a certain amount of exposure with a patchy cover of Juniperus and Betula. As shelter was available from SW winds a mosaic of Juniperus, Empetrum and open grassland developed with stands of tree birch. Juniperus an Empetrum declined as taller species arrived eventually shading out the dwarf shrub and heliophilous herbs. Together with the first patches of woodland it is likely that tall herb meadow flourished in wetter areas, but rapidly increasing values of Betula and Corylus indicate the imminent arrival of a closed woodland in which thermophilous arboreal species are also present.
<table>
<thead>
<tr>
<th>GA I</th>
<th>GA II</th>
<th>GA III</th>
<th>RPAZ</th>
</tr>
</thead>
<tbody>
<tr>
<td>GA I 14</td>
<td>Betula - Corylus</td>
<td>GA II 6</td>
<td>GAR5</td>
</tr>
<tr>
<td>GA I 3</td>
<td>Betula - Juniperus</td>
<td>GA II 5</td>
<td>Betula - Corylus</td>
</tr>
<tr>
<td>GA I 12</td>
<td>Empetrum</td>
<td>GA II 4</td>
<td>Betula - Juniperus</td>
</tr>
<tr>
<td>GA I 11</td>
<td>Poaceae - Betula</td>
<td>GA II 3</td>
<td>Poaceae - Cyperaceae</td>
</tr>
<tr>
<td></td>
<td>Rumex - Salix</td>
<td>GA II 2</td>
<td>Empetrum - Poaceae - Cyperaceae</td>
</tr>
<tr>
<td></td>
<td></td>
<td>GA II 1</td>
<td>Empetrum</td>
</tr>
<tr>
<td></td>
<td></td>
<td>GA III 1</td>
<td>Poaceae - Salix</td>
</tr>
<tr>
<td></td>
<td></td>
<td>GA III 2</td>
<td>Salix - Rumex - Poaceae</td>
</tr>
<tr>
<td></td>
<td></td>
<td>GA III 3</td>
<td>Empetrum - Poaceae</td>
</tr>
<tr>
<td></td>
<td></td>
<td>GA III 4</td>
<td>Empetrum - Poaceae - Cyperaceae</td>
</tr>
</tbody>
</table>

Local pollen assemblage zones and regional pollen assemblage zone for Glen Affric.
7.7 Prospects of a biostratigraphically determined deglacial chronology

Several studies have attempted to use pollen stratigraphy to establish a relative timescale for deglaciation (e.g. Pennington, 1978; McPherson, 1978; Lowe and Walker, 1981; 1985; 1991; Benn et al 1992). This rests on the principle that where deglaciation has been time-transgressive, the onset of sediment accumulation in enclosed lake basins would be progressively delayed. Hence the earliest Flandrian sediments should be found outside or near the ice limits with progressively younger deposits in basins upglacier. In terms of pollen stratigraphy, this pattern would be revealed by a complete early Flandrian pollen-stratigraphic sequence having been preserved in those sites where deglaciation occurred first, whereas the earliest pollen record would be absent from those sites where ice cover persisted for longer. Problems with regard to this methodology have been identified (Tipping, 1988; Lowe and Walker, 1990) and are addressed below. Whereas no evidence for sequential retreat has been identified at certain sites in Scotland (e.g. Tipping, 1984,1987; Lowe and Cairns, 1991) research in other areas such as Skye have revealed that pollen stratigraphy can be successfully employed to establish a relative chronology following the Loch Lomond Readvance maximum (Lowe and Walker, 1981; Walker and Lowe, 1985; Benn et al 1992).

The relationship between Loch Lomond Readvance limits and biostratigraphy in southern Ross-shire may be analysed on the basis of data presented in Chapters 5 and 8. The three pollen sites studied occupy different positions in relation to the Loch Lomond Readvance maximum limit in Glen Affric and to younger limits identified as relating to retreat stages. Geomorphological evidence suggests that the Glen Affric glacier declined actively by frontal retreat. Allowing for the assumptions outlined below, sequential retreat should be reflected in the three diagrams. In addition to detection of this in the pollen percentage diagrams, it was also postulated that it may be reflected in the deterioration diagrams.

A crucial consideration is to ensure that the earliest sediments are sampled and this was achieved by coring across the extent of the basin. The coring method involved preliminary sampling across the basin to ensure this was undertaken (cf. Section 7.2) to eradicate this problem. A related concern is, however, that the corer used was unable to sample basal sediments due to the metal tip, 15 cm long, below the sampling chamber being unable to penetrate the sediments. The result may have been that the base of the sampling chamber failed sample all organic material above the basal sediments. This may have occurred at site GAll but it was believed that the tip was below the sampleable material. A second problem is avoidable to some extent in that the selection of bedrock basins reduces the likelihood of the stratigraphic sequence having been disturbed by dead ice. The difficulties of stratigraphic disturbance identified by Gray (1975),
Macpherson (1978), Walker and Lowe (1979) and Lowe and Walker (1980) are more likely to be encountered in sedimentary basins which require dead ice in their formation; features such as kettle holes. Sites GAI and GAIL are located in rock basins but GAILI lies within an area of hummocky drift which may have contained dead ice.

Further difficulties associated with stratigraphy are, first, that the basal pollen assemblage zones may not be represented due to low water tables preventing sedimentation, as described by Walker and Lowe (1981). Tipping (1988) also addresses the problem of the basal sediments not having been sampled through minerogenic layers not having been sampled. It was obvious in the case of GAI that bedrock underlay the sediments sampled although for GAILI, a situation similar to that at Achnasheen where Lateglacial sediments were not sampled due to coarse minerogenic layers at depth (Sissons, 1982) and at other sites (Sissons, 1980) was thought to apply. It was evident, however, that the corer penetrated some depth into minerogenic sediments and it is considered that the earliest Flandrian sediments were probably recovered.

A fundamental problem concerns the question of whether the sequence of pollen stratigraphic changes is discernible and consistent between sites (Tipping, 1988). Varying sedimentation rates may be problematic here, together with the coarseness of the sediment and low productivity of the early pioneer plant communities. The overriding influence of taxa which are unrelated to major taxa of the vegetation succession is also considered to be a problem. Sedimentation rates are considered here to be sufficiently similar and not to be a significant problem. (Taxa do not tend to peak synchronously).

A final difficulty applicable to this study is the assumption that ice decay is predominantly by frontal retreat rather than by in situ decay. Geomorphological evidence from Glen Affric demonstrates that frontal retreat of the ice margin occurred in at least the lower parts of Glen Affric. This would suggest that the effects of stagnating ice, delaying sedimentation should be minimal. The depth of ice may, however, have affected the pollen stratigraphy, in terms of the lowest pollen assemblages recorded.

The number of zones present does not steadily decrease with proximity to ice source as revealed in other studies (Walker and Lowe, 1989; Benn et al 1992). Such a trend reflects the greater timescale over which pollen could accumulate at sites which were first and which are most distant from the ice source areas which tended to become free of ice last. Some evidence for such a pattern is evident in so far as the outermost profile reveals the earliest sediments as demonstrated by pollen assemblages and a regional comparison. The presence of a Rumex peak is recorded at the site furthest upvalley as well as in the peripheral site. Its context within the regional pollen spectra is of significance in that it is clear that the innermost site was free of ice prior to the immigration of juniper to the area. Its expansion occurred quite some time after sedimentation commenced at GAI. This is
in accordance with the proposed period of active retreat under continuing cold conditions. The evidence in GAI11 for a slight climatic revertence supports this assertion.

The evidence suggests that basal Flandrian pollen assemblages in Glen Affric reflect the operation of three factors: first, local vegetation colonisation; secondly, regional plant succession and thirdly, variation in the timing of deglaciation.

Comparing these results with other data related to deglaciation, as in Skye, most of the main icefield had melted by the time that juniper was colonising the island as sites closest to the ice sources show rising *Juniperus* percentages at the base. (On the basis of C14 dates for the early Flandrian expansion of juniper in the Inner Hebrides it is likely to have occurred no later than 9.6 ka BP.) A pattern of initial frontal retreat followed by in situ decline of glaciers in Skye has been identified in both pollen stratigraphic and geomorphological analyses. This situation is also indicated in southern Ross-shire, although sites near the ice source have not yet been investigated. In the Rannoch Moor area, dates for deglaciation vary, possibly due to spatial variations in the pattern of deglaciation and the progressive retreat of the ice front but the pattern of deglaciation remains less clear.

The inferences that can be made based on biostratigraphic analyses in Glen Affric are limited due to the small number of sites investigated, the local vegetation differences at the sites, the differences in the nature of the sites themselves and the possibility of methodological problems with regard to sampling the earliest sediments. It would be necessary to undertake future work in the field area to attempt to establish the degree of synchronicity between deglaciation in adjacent valleys. Tipping's (1984, 1988) methodology of selecting pairs of basins, allowing the identification of closely comparable pollen stratigraphies could be usefully employed here to isolate more successfully the regional pollen component from the local one.

### 7.8 Regional comparisons and correlations

Pollen studies have revealed a characteristic sequence of Pollen Assemblage Zones common to most Flandrian pollen profiles for Scotland (Walker and Lowe, 1991). The diagrams for Glen Affric broadly conform to this. Elsewhere, pollen diagrams exhibit recurrent features indicative of an orderly vegetation succession in response to regional climatic change. (The influence of the ameliorating climate is discussed in Chapter Nine.) This inferred succession consists of biozones characterised by successive maxima in pollen of Poaceae, *Rumex, Empetrum, Juniperus, Betula* and *Corylus*. Figure 7.13 shows the regional pollen assemblage zones for Skye, Wester Ross, northern Scotland, western Scotland and eastern Scotland. The vegetational sequence that has been inferred
Figure 7.13
Correlation of regional pollen assemblage zones

<table>
<thead>
<tr>
<th>Regional pollen zone</th>
<th>Regional pollen zone</th>
<th>Regional pollen zone</th>
<th>Regional pollen zone</th>
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</thead>
<tbody>
<tr>
<td>Western Skye</td>
<td>Eastern Skye</td>
<td>Glen Affric</td>
<td>Rannoch Moor</td>
</tr>
<tr>
<td>SWSK 7 Betula - Corylus -</td>
<td>ESK 11 Betula - Corylus</td>
<td>GAR 5 Betula - Corylus</td>
<td>Betula - Pinus</td>
</tr>
<tr>
<td>Poaceae - Rumex</td>
<td>ESK 10 Betula</td>
<td>Betula - Corylus</td>
<td>Betula - Corylus - Ericaceae</td>
</tr>
<tr>
<td>SWSK 6 Betula - Empetrum</td>
<td>ESK 9 Betula - Juniperus</td>
<td>GAR 4 Betula - Juniperus</td>
<td>Betula - Juniperus</td>
</tr>
<tr>
<td>Poaceae - Rumex (Juniperus)</td>
<td>ESK 8 Juniperus - Poaceae</td>
<td>GAR 3 Poaceae - Cyperaceae (Empetrum)</td>
<td>Juniperus - Betula</td>
</tr>
<tr>
<td>SWSK 5 Artemisia - Rumex</td>
<td>ESK 7 Empetrum - Poaceae</td>
<td>GAR 2 Empetrum - Poaceae</td>
<td>Empetrum</td>
</tr>
<tr>
<td>Cyperaceae - (Empetrum)</td>
<td>ESK 6 Cyperaceae - Rumex</td>
<td>GAR 1 Poaceae - Salix - Rumex</td>
<td></td>
</tr>
<tr>
<td></td>
<td>ESK 5 Rumex - Artemisia</td>
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7.8.1. Comparison of trends in spectra and pollen stratigraphic events.

Several early Flandrian pollen stratigraphic events, occurring in pollen stratigraphies from sites in northern and western Scotland have been identified. These include an initial Rumex peak, an Empetrum rise, revertence and maximum, a Juniperus maximum, a Betula - Juniperus phase and finally a Betula - Corylus stage. The presence or absence and importance of these events in Glen Affric is assessed.

A Rumex peak is identified in GAI 1 and GAIII 1 and 2. Values for Rumex pollen rise to a maximum in both GAI and GAIII. This contrasts with those for sites elsewhere where initial Rumex values are high such as in Skye at Glen Arroch, Sligachan 1 and 2 (Benn et al 1992). The importance of the initial stages in relation to deglaciation are discussed below. Whilst sites peripheral to Loch Lomond Stadial glaciers on Skye have Rumex through to the Empetrum-Juniperus stages, Affric values are lower. However, trends although less impressive may be as significant. The Rumex peak may be equivalent to the base of Pennington's (1978) Poaceae - Rumex - Empetrum zone.
In many early Flandrian pollen diagrams from Scotland *Empetrum* is seen to expand following *Rumex* (Pennington, 1977, 1978; Lowe and Walker, 1978, 1980, 1981; Walker and Lowe, 1977, 1979, 1980, 1985) and this pattern is also reflected in studies of pioneer plant colonisation (Matthews, 1978). A minor peak in *Empetrum* values in GAI 1 is followed by a revertence registered in GAI 2 prior to its marked expansion. This pattern is not reflected in the lower zones for either GA I or GAI. An *Empetrum* revertence episode is evident at Glen Arroch in Skye (Benn *et al*, 1992) where it is accompanied by similar rises and falls in *Betula* and *Juniperus*. In GAI, *Betula* values fall immediately prior to those of *Empetrum* but juniper has not yet established itself in the Glen Affric area. The evidence for a *Empetrum* revertence episode also reported from other sites involves limited disruption to plant succession during the early stages of vegetation colonisation. The phenomenon is interpreted as a result of minor climatic cooling (Chapter 8). The taxa peak in the correct order compared with the general sequence described above.

A distinct *Empetrum* maximum is observed in all the Glen Affric sites. Whereas there is a distinct rise to a substantial maximum in GAI 2 and GAI 3, *Empetrum* values for GAI 2 are initially maximal, at lower values and at once decreasing. In GAI these are accompanied by high values for *Huperzia selago*, a situation which is mirrored at Luib 1,2 and 3 in Skye (Benn *et al*, 1992). In comparison with other sites *Juniperus* is poorly represented. Like Glen Affric, however, a number of sites on Skye exhibit a poor representation of *Juniperus* although this is related to exposure to south westerly winds. An increase in *Juniperus* is preceded by high values of *Myriophyllum alterniflorum* at GA I and GAI 2. This pattern is recorded at a number of sites on Skye (Walker and Lowe, 1990) and sites on Mull (Lowe and Walker, 1986b, 1987). Its increase is reported to be accompanied by taxa indicative of shallow water environments with taxa such as *Potamogeton, Equisetum* and *Menyanthes*. Of these, *Equisetum* is the only taxon to be similarly represented in GAI and GAI 2. Although it is suggested that such fluctuations in *Myriophyllum* could reflect local variations in water quality, it is likely that regional climatic forcing is responsible, as invoked to explain the changes on Skye (Benn *et al* 1992).

The *Betula* or *Betula - Juniperus* phase is fairly pronounced, as in eastern Skye. The curves for *Betula* and juniper for south-west Skye are less significant. A similar trend occurs for the *Betula - Corylus* phase. The immigration of other trees with the arrival of *Corylus* is reported at some sites. There is negligible *Alnus* in Glen Affric cores, as is the case for Glasscnock (Robinson, 1977).

A number of apparent anomalies occur in the profiles. The high values of Cyperaceae in GAI 2 that is considered to be due to redeposition may be present elsewhere for this
Salix is also relatively important in southern Ross-shire. It exists in wetter areas and is associated with Poaceae, Cyperaceae, Filipendula, and Rumex but probably reflects local conditions.

7.8.2 Comparison of pollen profiles, assemblage zones, vegetation development between Glen Affric and other areas in Scotland.

A tentative correlation between RPAZs is attempted. General similarities exist between all regions (Figure 7.13). The notable differences between the findings for the present area and other areas are discussed below.

With regard to the western Highlands there is no evidence for a pioneer herbaceous phase on Rannoch Moor preceding the Empetrum zone as described in GAR (although there are significant percentages of Rumex/Oxyria in basal horizons). No rise in Empetrum to a distinct postglacial maximum such as that recorded in the present study. It has been proposed that these initial high values of Empetrum with no pioneer phase are due to First, delayed sedimentation. Secondly, impoverished natures of the pioneer pollen rain. Thirdly, very rapid rate of immigration of Empetrum and associated dwarf shrub heath plants. The early Empetrum phase is followed by a distinct phase of Juniperus dominance of greater proportions than at Glen Affric.

In profiles for the eastern Highlands, an initial herb phase of herb - dominated communities is missing from the basal sections of profiles from sites such as Mollands and Cambusbeg in the Teith valley. A herb dominated pollen assemblage is, however, present at Tynaspirit. The early shrub dominated vegetation that developed in which Juniperus and Empetrum were the main components together with lesser but still significant amounts of Salix is similar to that for Glen Affric. The relatively high importance of Juniperus and Salix through the Betula phase is likewise registered in the present study sites as is the rapid expansion of Corylus in the area through the established birchwood indicated in the profile following the Betula maximum.

The pollen analytical work carried out on Skye demonstrates a number of differences: first it appears that juniper was quicker to reach Skye. The work carried out by Walker and Lowe (1990) revealed that the variation in exposure to prevailing winds was the dominant factor influencing the development of vegetation patterns, which do vary across the island.

Pollen diagrams, pollen assemblage zones and inferred vegetation development may be seen to vary considerably from E to W. Southern Ross-shire is situated in a fairly critical position between sites reported both to the west and east. Glen Affric itself lies in
the east of the field area, and although associated with the western mountains is reasonably close to the east coast. It is useful to consider whether the information revealed for southern Ross-shire conforms more to the east or to the west or has intermediate tendencies. Several points can be made: first, the diagrams for Glen Affric would may be expected to contain more pine, as found in eastern Highlands; second, *Betula* and *Corylus* important in the north and west and this is certainly the case.

The boundaries of RPAZs provide a basis for time-stratigraphic correlation across the area. In the absence of radiocarbon dates for the three sites analysed, dates from adjacent studies may be used tentatively to date the juniper rise and other pollen events. The rise and fall in *Empetrum* at Varagill I on Skye is dated to 10,220+/- 150 yr BP (Walker and Lowe, 1990).

Finally, the environmental and climatic implications of the pollen investigations may be addressed. The questions arise as to whether the successive peaks in *Empetrum* and *Juniperus* reflect a sustained rise in Flandrian temperatures. It may be argued that increased *Juniperus* preceded by high counts of *Myriophyllum* indicative of regional climatic forcing. The high counts of deteriorated pollen may also indicate severe paraglacial activity in the early postglacial.

### 7.9 Conclusions

Three pollen sites have been investigated and information yielded chiefly in terms of early Flandrian vegetation development. These data have assisted in the interpretation of the nature of succession following the deglaciation of a local ice mass in Glen Affric. The high resolution of the sampling has enabled the nature of immigration to be considered in much detail. The sites are important for determining the local pattern of early Flandrian vegetation development for the area, for which there was previously no information. They also provide the possibility of adding to and clarifying the detail of the regional picture of early Flandrian vegetation history. An early Flandrian revertence phase is present, adding support to the evidence of its occurrence over a wide region of western Scotland. It has been possible to elucidate east-west differences in inferred vegetation development across Scotland. The study has considered the use and efficacy of pollen preservation analysis and has concluded that such activities are important in the interpretation of pollen taxonomy.
CHAPTER EIGHT

Conclusions

8.1 Conclusions

The lateral and vertical extent of two, late Quaternary ice masses have been determined in southern Ross-shire. These periods of ice presence are assigned to the Loch Lomond Readvance and the Late Devensian Ice Sheet. The ice masses had sources on the high ground in the field area and, in the case of the former, terminated within southern Ross-shire whereas the latter ice sheet extended onto the continental shelf in the west and eastwards into the North Sea.

8.2 Loch Lomond Re-advance

The extent of the Loch Lomond Readvance glaciers in southern Ross-shire is clearly delimited by geomorphological evidence.

Thirteen glaciers flowing east and west from the central mountainous area have been identified and their lateral and vertical extents determined. Although glaciers assigned to the Loch Lomond Readvance have previously been mapped in the area (Peacock, 1970, Sissons, 1976, Bennett, 1991), this thesis is a more comprehensive study which adds significant detail to our knowledge of the northern limits of the Loch Lomond Readvance Ice Sheet in Scotland. Geomorphological evidence shows that, at its maximum extent, the ice sheet contained over 129 cubic kilometres of ice with a central elevation of over 800 metres and a thickness of y metres. It is noteworthy that this is over 100m higher than that predicted by Bennett (1991) on the basis of limits inferred from lateral and end moraines. The central ice sheet was fed by numerous corrie glaciers and the discharge through the valley glaciers reached the sea in the west but did not extend below z metres in the east.

The development of large valley glaciers in southern Ross-shire was enabled by the low mean annual sea-level temperatures of c. 5 °C, calculated from the geomorphological reconstructions and furthered by pollen analytical inferences. The pattern of ELAs calculated for the reconstructed glaciers range from 460 m in the NW to 670 m in the SE. This implies that the main snow-bearing winds were from the west, with the amount of snow decreasing gradually to the east.
The mapped evidence, together with palaeoclimatic reconstructions, suggest glaciological conditions of highly active glaciers with relatively high velocities and discharge. The lower elevation of the glacier snouts in the west relates to greater glacier activity in this sector of the ice sheet and corresponds with the most deeply dissected mountains, deep troughs and numerous corries. Lower glacier activity is indicated in the NE where there was less snow and therefore less ice and lower gradients.

The present research is asserted as being more accurate than previous studies of the Loch Lomond Readvance in the area. The maximum extent of the southern tongue of the Affric Glacier corresponds with that suggested by Sissons (1976) but conflicts with a more extensive limit proposed more recently by Bennett (1991) by several kilometres. The maximum limits given by Sissons (1982) for Glen Doe have been confirmed in part but the mapping undertaken in this study proposes a more extensive corrie glacier in the valley of Allt Coire Sgreurnh (NH1514) and defines the extent of a tongue of ice that spilled over a col at NH158181 and possibly at NH153167 into Gleann Fada NH1617. The limits for the maximum extent of this local glaciation are, throughout the area, more extensive than those postulated by the Geological Survey (1913a) and by Sissons (1982). The present research has indicated that the limits at Achnasheen are in fact associated with the latest local glaciation and not a previous readvance as proposed by Sissons.

At the northern margin of the present field area, the mapped Loch Lomond Readvance limit confirms, to some extent, that of Robinson (1977). Several refinements to previously given limits in Strathcarron are proposed, together with the extension of the maximum lateral limit to Achnasheen. It is suggested that the limits at NH 013525 and NH 025487 (Robinson, 1977) are well within the maximal limit and may relate to deglaciation (Section 5.7). The lateral limit defined at the south-easternmost part of the field area is similar to that identified by Sissons (1977) and to some extent Peacock (1970), although the latter failed to pinpoint an actual limit.

8.3 Late Devensian ice sheet

Mapping indicates that the southern part of the field area was almost certainly covered by the last ice sheet. The low gradient of the ice sheet towards the north also suggests that the northern area of southern Ross-shire was covered by ice during this period. Such a conclusion seems at variance with the presence of periglacial detritus and more abundant periglacial landforms on mountains in the north of the field area. However, this evidence, because of its inconsistent lower altitude cannot be interpreted in terms of an ice sheet trimline. It is considered to have survived Late Devensian glacial activity due to the thermal regime of the glacier ice because of differences in topography and ice thickness.
Evidence is present at the margins of the field area for a readvance associated with the last ice sheet, but predating the Loch Lomond Readvance. End and lateral moraines and drift limits indicate that there was a readvance, possibly relating to the Wester Ross Readvance, that reached maximum positions a few kilometres downvalley of the identified Loch Lomond Readvance limits in Glen Meinich, Glen Elchaig and Glen Attadale.

8.4 Holocene Vegetation development

Three pollen sites have been investigated which have assisted in the interpretation of the nature of succession following the deglaciation of a local ice mass in Glen Affric. An early Flandrian revertence phase is present, adding support to the evidence of its occurrence over a wide region of western Scotland. It has been possible to elucidate east-west differences in inferred vegetation development across Scotland. The study has considered the use and efficacy of pollen preservation analysis and has concluded that such activities are important in the interpretation of pollen taphonomy.

8.5 Geomorphological Mapping

This rigorous assessment of geomorphological evidence to determine the lateral and vertical extents of ice masses forms an integral and important part of this thesis. The methods used for reconstructing former glaciers, on the basis of geomorphological mapping of glacial and periglacial evidence, such as initiated by Sissons (e.g. 1979d) and used by other researchers (e.g. Ballantyne, 1989b) is supported by the results presented in this study. However, the conclusions drawn from some of these earlier studies is weakened because they have not mapped all glacial features together with associated periglacial evidence. This is necessary in order to establish the three dimensional form of former ice masses.

The 'trimline' technique is an important method whereby it may be possible to establish the former upper limits of ice sheets. This study has tested the procedure for mapping both erosional and depositional trimlines, especially in the accumulation areas of the former Loch Lomond Readvance ice masses, in southern Ross-shire but also in relation to the Late Devensian ice sheet. In particular, the identification and mapping of depositional trimlines, as well as 'erosional' trimlines has made it possible to be much more confident in delimiting the vertical extent of former glaciers in southern Ross-shire.

Several limitations to the use of trimline data are noted as a result of this study. Firstly, the data are more readily applicable to larger masses of ice, including valley glaciers of the size depicted for this study, than for smaller, often discrete, glaciers such as corrie glaciers.
Secondly, an exact numerical value cannot always be ascertained for the height of a trimline and the altitude of the former ice mass has to be estimated or inferred to a degree in some cases. Thirdly, variations in lithology and structure can mean that the interpretation of evidence below and above the trimlines becomes difficult, due to the fact that rocks with differing characteristics vary significantly in their response to both glacial and periglacial processes. Fourthly, for investigations of Loch Lomond Stadial glaciers, it is known that below about 500m periglacial weathering processes were less effective than at higher elevation and accordingly produced little trimline evidence.

8.6 Wider implications

The results of this research can now be reviewed in relation to previous work in the wider context of the emerging national and global view of Late Quaternary environments. Detailed geomorphological mapping of depositional landforms has revealed evidence for widespread active retreat of Loch Lomond Readvance glaciers with some localised stagnation. This is interpreted to reflect an initial post-maximum period of ice margin fluctuation followed by a period of local in situ ice stagnation. This is consistent with the pollen record for the area which reveals a possible climatic revertence which may be associated with the glacier readvances or stillstands that have been identified in the form of recessional moraines. The rapid thermal amelioration following the period of climatic fluctuation accompanied by recessional moraine formation hypothesised by Benn et al. (1992) is also consistent with the data presented here. A rapid climatic amelioration is also suggested in evidence from an ice core from Greenland (GISP2), as oxygen isotope data suggest that the Younger Dryas ended abruptly over a period of 50 years in response to a trigger in the North Atlantic climate system.

The surface of the Late Devensian ice sheet over southern Ross-shire has been shown to have been at least 1100 m altitude. This is consistent with recent evidence for an ice sheet up to 1500 m recently suggested by Somme (1995). It is probably also consistent with the reconstructions given by Ballantyne for Skye (1990), although the evidence for Wester Ross (Reed, 1988) appears to conflict with the evidence presented here.

8.7 Future research.

There is a need to obtain a firm age for the proposed transection glacier complex, inferred to be of Loch Lomond Stadial age. Radiometric dating of basal organic sediments within the proposed limits would provide a limiting date for deglaciation. Confirmation of the proposed limits could be achieved through further stratigraphic analyses beyond the proposed limits:
sites which show a clear tripartite sequence representing the Lateglacial interstadial and Loch Lomond Stadial and Holocene deposits would confirm non-glaciation during the Loch Lomond Stadial.

Glacial evidence believed to relate to a readvance associated with the deglaciation of the ice sheet and predating the Loch Lomond Readvance has been located in a number of the mapped valleys. Other ice-marginal deposits were observed during fieldwork beyond the extent of mapping in positions downvalley of the Loch Lomond Readvance. These may correlate with those occurring within the defined field area and further mapping may elucidate this relationship. Further mapping of deposits is also required in areas such as upper Glen Ling and Glen Strathfarrar where the nature of deglaciation has not been fully ascertained.

This study focuses on morphological analysis of the glacial and periglacial evidence. Detailed sedimentological analyses of the glacial tills may yield valuable information regarding till provenance, genesis and depositional environment. Further sedimentological analyses may also provide greater insights into the former glacier regimes for the Loch Lomond readvance and Late Devensian ice sheet. Similarly, it may be possible to ascertain the relative or even absolute age and genesis of the periglacially weathered material by further analysis of the constituent particles and by analysis of mountain-top soils.

There is also scope for further analysis of the pattern of deglaciation of the Loch Lomond Readvance glaciers, particularly relating to the correlation of individual glaciers. This would ideally be undertaken in association with further pollen stratigraphic analyses and the dating of sediments from sites for each valley to provide an absolute chronology of deglaciation.

The present study proposes that the limits at Achnasheen are of Loch Lomond Readvance origin. This implies that the limit in the valley immediately to the north is contemporaneous. If this is the case it is probable that the limits of the Loch Lomond Readvance given by Robinson (1977) are underestimated and that these represent various stages in the deglaciation of the ice from its maximum at Achnasheen. Such an assertion is consistent with the interpretation of the extent and pattern of glaciation for the adjacent Glen Carron and Strathcarron given in this study. Further detailed mapping is required in the area NW of Achnasheen. Similarly, the area immediately to the south of the field area also remains to be mapped in detail. Such a study would enable the pattern of glaciation depicted in the present study to be placed more accurately in the context Late Quaternary glaciation in Scotland in general.
Recent research on Skye (Dix, Pers. Comm; Lowe, 1993) has explored the possibilities of identifying subaqueous morainic ridges. The precise location of Loch Lomond Readvance limits in Loch Duich and in Loch Long are uncertain. It is considered that seismic analyses of the floors of these lochs in the vicinity of proposed limits may be useful, in addition to identifying ice marginal positions for still-stands during deglaciation. Similarly, the technique may be further employed (as in Lowe, 1993) with respect to the various lochs in the eastern valleys such as Loch Beanacharain.

The present study sought to define the nature of the shoreline evidence relating to the glaciers in the field area. It has been possible to map and survey only a limited portion of the former coastline of southern Ross-shire. A detailed analysis of the coastal area of the whole of the Kyle of Lochalsh northwards to Achmore is still required. This may clarify the positions of the Loch Lomond Readvance and previous (Wester Ross Readvances) by virtue of the associated shorelines expected.

Finally, little has been mentioned with regard to Holocene landform development. There is considerable scope within the field area for surveying paraglacial features, landslides, both inside and outside the defined glacial limits of which there are many, and analysis of the many alluvial landforms. It may be possible in particular to relate the various landslides to glacial limits or stages of deglaciation.
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Figure 5.2
Figure 5.30

Area above or beyond ice (contours at 50m intervals)

Contoured glacier surface at 50m intervals
Figure 5.29

Area above or beyond ice (contours at 50m intervals)

Contoured glacier surface at 50m intervals
Figure 5.32

Area above or beyond ice (contours at 50m intervals)

Contoured glacier surface at 50m intervals
Figure 7.9I - III

I Total deteriorated pollen as a % of total land pollen

II Representation of each class as a percentage of total recognisable deteriorated pollen

III Inner curve - Percentage of Total Recognisable Deteriorated Pollen
Outer curve - Percentage of Total deteriorated Pollen
Figure 7.11 III

Well preserved Corroded Broken Crumpled/folded Degraded

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Betula Corylus Salix Juniperus Empetrum Calluna Poaceae Cyperaceae

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GA III 6
GA III 5
GA III 4
GA III 3
GA III 2
GA III 1
Figure 7.11 IV

BETULA

CORYLUS

SALIX

JUNIPERUS

EMPETRUM

CALLUNA

POACEAE

CYPERACEAE