Chapter 2 The Quaternary in Scotland

D.G. Sutherland and J.E. Gordon

Pre-glacial landform inheritance and the effects of glacial erosion

The broad outlines of the Scottish landscape owe their origins to a combination of geological and tectonic controls and geomorphological processes in pre-Quaternary time. However, there are few on-land deposits that pre-date the Quaternary and the great majority of the relict features in the landscape are erosional and therefore difficult to date. Consequently there has been a range of interpretations and reconstructions concerning the existence, ages and origins of erosion surfaces, drainage systems and weathered bedrock (Linton, 1951b; Walton, 1964; George, 1965, 1966; Godard, 1965; Sissons, 1967a, 1976b; Haynes, 1983; Hall, 1991). Nevertheless, detailed regional studies, together with consideration of the sediments deposited on the adjacent continental shelves, have allowed progress to be made in recent years (Hall, 1983, 1986, 1991; Le Coeur, 1988) in understanding the longer-term evolution of the present landscape and its large-scale features.

Study of deep-sea cores recovered from the floor of the North Atlantic has suggested that the first mid-latitude ice-sheet glaciation during the present sequence of ice ages occurred around 2.4 Ma (Shackleton *et al.,* 1984; Loubere and Moss, 1986). Although no direct evidence has been found for glaciation of Scotland or the neighbouring shelves at that time, the relatively minor decline in temperature that would be necessary for glaciers to develop in Scotland suggests glaciation is likely to have occurred at least in the Highlands.

The general character of the topography on which the earliest glaciers developed may not have been too different from that of today, with a mountainous western Highland zone with a range of relief not dissimilar to that of the present and fringing lowlands, particularly on the eastern side of the country. The principal watershed was likely to have been located in the west, with major rivers draining to eastern estuaries. The North Sea was probably smaller than today, and the low-lying eastern coastal fringe consequently wider. The Orkney Islands were probably then joined to the mainland, but the western island groups may have been broadly the same as at present, with the exception that those islands close to the mainland (Skye, Mull, Scarba–Jura– Islay), which owe their isolation to glacial erosion, would then have been peninsulas. These inferences can be made from the known distribution of pre-glacial weathered bedrock profiles and erosion surfaces (Linton, 1959; Godard, 1965; Sissons, 1967a, 1976b; Hall 1985, 1986, 1987), the occurrence of Neogene sediments on neighbouring continental shelves (Andrews *et al.,* 1990), and consideration of the tectonic and erosional history of the Highland block during the Mesozoic and Tertiary (Watson, 1985; Hall, 1991).

Although areas of weathered bedrock occur locally in the west (Godard, 1961, 1965; Le Coeur, 1989), pre-glacial relics are most abundant in the Buchan area of north-east Scotland, where they exist in most notable form in deposits of fluvial and/or marine gravels at an altitude of around 150 m OD (see Moss of Cruden and Windy Hills) (Flett and Read, 1921; Koppi and Fitzpatrick, 1980; McMillan and Merritt, 1980; Kesel and Gemmell, 1981; Hall, 1982, 1984c) and in frequent occurrences of chemically decomposed bedrock (see Hill of Longhaven, Clunas and Pittodrie) (Phemister and Simpson, 1949; FitzPatrick, 1963; Hall, 1984b, 1985, 1986, 1987; Hall and Mellor, 1988; Hall *et al.,* 1989a). These features are of considerable interest, recording aspects of the Neogene geological history of Scotland. They occur in what was probably the most extensive on-land area of north-west Europe where such old features have been preserved despite a history of glaciation. Additionally, the distribution of the decomposed bedrock outcrops provides an indication of the intensity of glacial erosion in north-east Scotland (Hall and Sugden, 1987; Sugden, 1989) and illustrates graphically the localized nature of this process, even though certain of the weathered bedrock profiles may be truncated (cf. Godard, 1989).

Elsewhere in Scotland the effects of glacial erosion give a distinct aspect to the landscape, albeit one that varies from one part of the country to another (Figure 2.1) and (Figure 2.2). In the eastern Highlands, the landforms of glacial erosion, whether corries fretting the edges of the major mountain plateaux (see the Cairngorms and Lochnagar), or spectacular through valleys such as the Lairig Ghru or Glen Tilt, are clearly superimposed on a pre-existing topography, parts of which have survived in a largely unmodified form (Figure 2.3). In such areas glacial erosion has operated selectively, although to striking effect, in creating glacial troughs, breaches and diversions of drainage (Linton, 1951a, 1963; Sugden, 1968). In the western Highlands and adjacent mountainous islands, however, erosion has been much more intense (see the Cuillin), as is shown by the frequency of corries, sharp ridges, pyramidal summits, overdeepened valleys, rock basins and fjords which often dominate the scenery (Linton, 1951a, 1957, 1959, 1963; Linton and Moisley, 1960; Sissons, 1967a, 1976b; Clayton, 1974; Haynes, 1977a; Dale, 1981). The lowlands, too, have received an erosional imprint from glaciation, whether in the form of knock- and-lochan topography (areal scouring) so typical of the Lewisian gneiss terrain of north-west Sutherland (Figure 2.4) and the Outer Hebrides (Linton, 1959; Godard, 1965; Gordon, 1981) or as the streamlined hills and crags and tails of the Midland Valley (Linton, 1962; Burke, 1969).

Zone	**Lowlands**	**Uplands**
0	No erosion Head on weathered rocks and slopes Outwash in concavities Rare occurrences of till on weathered rock	No erosion Outwash on valley floors Solifluction deposits on slopes Boulder fields and tors on divides
I	Ice erosion confined to detailed or subordinate modifications Concavities drift mantled but convexities may show some ice moulding Occasional roches moutonnees Ice-scoured bluffs in favourable locations	Ice erosion confined to detailed or subordinate modifications Suitable valley slopes ice steepened Entrenched meanders and spurs converted to rock knobs Interfluves still commonly Zone 0
II	Extensive excavation along main flowlines so that concavities may be drift free or floored by outwash or post-glacial deposits. Isolated obstacles may be given ovoid or cutwater forms if of soft rock, or crag-and-tail with associated scour troughs if hard Margins of larger masses converted to ice-scoured bluffs or planar slopes	Conversion of pre-glacial valleys to troughs common, but usually confined to those of direction concordant with ice flow Some diffluence; transfluence rare Interfluves may be Zone I or even Zone 0, and separated from troughs by well-marked shoulders
III	Pre-glacial forms no longer recognizable but replaced by tapered or bridge interfluves with planar slopes on soft rocks, and by rock drumlins and knock-and-lochan topography on hard rocks	Transformation of valleys to troughs comprehensive giving compartmented relief with isolated plateau or mountain blocks. Zone 0 may still persist on interfluves at sufficiently great heights Ice moulding extends to high summits
IV	Complete domination of streamlined flow forms even over structural influences	Upland surfaces given knock-and-lochan topography (sometimes of great amplitude at lower levels) Lower divides extensively pared or streamlined

It has been argued (Sutherland, 1984a) that as the ice-flow directions at the maximum of the last ice-sheet glaciation (during the Late Devensian, see below) were, in places, at a high angle to the direction of ice movement that was responsible for the formation of major erosional landforms, then those features must have originated during earlier glaciations. It is most probable that in Scotland there were numerous periods of either partial or ice-sheet glaciation during the Quaternary. During the former, erosional forms such as corries would have been progressively fashioned, and during the latter, major glacial breaches, overdeepened valleys and lowland ice-moulding would have formed. It may be further argued that as there would very probably have been extensive areas of deeply weathered bedrock throughout Scotland at the onset of glaciation, the initial glacial phases were likely to have had the greatest erosional impact, with later glaciations chiefly inheriting and only slightly modifying erosional landforms.

Two principal factors external to the relief and geological structure have influenced the distribution and degree of development of landforms of glacial erosion in Scotland. Major features (such as glacial breaches, overdeepened valleys and zones of areal scouring; (Figure 2.1) and (Figure 2.2) were controlled principally by the dynamics and thermal regimes of the ice masses that were responsible for their development (cf. Sugden and John, 1976). They show regional patterns as a consequence (Linton, 1959; Clayton, 1974; Boulton *et al.,* 1977; Haynes, 1977a; Gordon, 1979; Hall and Sugden, 1987), not only with respect to the major ice-sheet that centred on the western Highlands but also around the minor ice centres such as that on the Outer Hebrides. In contrast, small features such as corries, which formed during periods of partial glaciation, have regional distributions and characteristics more closely related to palaeoclimate (Linton, 1959; Sissons, 1967a; Sale, 1970; Robinson *et al.,* 1971; Dale, 1981) and only secondarily or locally to pre-existing relief and geological controls (Thompson, 1950; Haynes, 1968; Sugden, 1969; Gordon, 1977).

Quaternary events prior to the Late Devensian ice-sheet glaciation

Deposits attributed to the Late Devensian ice-sheet glaciation can be recognized throughout most of Scotland and hence form a useful datum. Prior to this event, the onshore Quaternary record is fragmentary and, as a consequence, correlation between sites with evidence of earlier events is difficult. Offshore, particularly in the North Sea Basin, much longer sequences of Quaternary deposits are known and these provide invaluable information for interpreting the onshore record. Even offshore, however, there are major unconformities and difficulties of correlation.

Marine deposits and landforms

Pre-Late Devensian marine deposits and landforms are known from localities around almost all the Scottish coasts (Figure 2.5). Possibly the earliest marine event for which there is evidence onshore is represented by the ice-transported mollusc shell fragments in the Kippet Hills area of Aberdeenshire. These shell fragments are contained within glacial deposits of presumed Late Devensian age, but their derivation from an Early Pleistocene deposit, originally suggested on faunal grounds by Jamieson (1882a), has been confirmed by amino acid analysis (D. G. Sutherland, unpublished data). The shell fragments are probably derived from marine deposits close to the Aberdeenshire coast which correlate with the Aberdeen Ground Formation of the North Sea sequence (Stoker *et al.,* 1985).

The Aberdeen Ground Formation accumulated during the whole of the Early Pleistocene and the early Middle Pleistocene and the greater part of its thickness is comprised of silts, clays and fine sands deposited in a relatively shallow (around 50 m) sea as part of a prograding delta sequence (Stoker and Bent, 1987). Along the western margin of the central North Sea Basin, however, a number of beds of coarser sands and gravelly sands have been found in the Early Pleistocene sediments and these were deposited as channel lags in a nearshore environment with water depths of less than 15 m. These sediments are now at depths greater than 150 m below present sea level, which may in part be due to basin subsidence. However, such subsidence would also apply to the deeper-water sediments with which they are interbedded and hence the nearshore sediments indicate low relative sea levels during the Early Pleistocene which seem most probably to be the consequence of glacio-eustatic lowering of sea level. The uppermost of these beds occurs between the top of the Jaramillo palaeomagnetic event (*c.* 990 ka) and the Brunhes–Matuyama palaeomagnetic boundary (780 ka) and may broadly correlate with the first evidence found for glacial sedimentation in the central North Sea by Sejrup *et al.* (1987). Such a correlation would support the attribution of the other beds of nearshore sediment in these Early Pleistocene deposits to glaciations. For much of the Early Pleistocene, however, no glacial sediments have been recognized in the North Sea basin. Apart from the low sea levels mentioned above, the approximate position of the Scottish coastline during this period cannot be ascertained because of the erosion of the western margin of the Aberdeen Ground Formation during Middle and Late Pleistocene glaciations. Certain of the wide erosional platforms described between altitudes of −30 m to −70 m off the east coast (Stoker and Graham, 1985) may have been formed at this time.

Onshore, a distinct set of marine deposits are the so-called 'high-level' shell beds which occur at altitudes of up to 150 m OD (Figure 2.5). The genesis of these deposits has been debated since the discovery of the first sites in the last century (see Clava, Tangy Glen and Afton Lodge) (Bell, 1893a, 1893b, 1895a; Horne *et al.,* 1894, 1897; Eyles *et al.,* 1949; Holden, 1977a; Sutherland, 1981a; Merritt, 1990b). Their origin and chronology are poorly understood, but whether they are *in situ* or are large, ice-transported erratics, they clearly represent marine invasion of the Scottish coastal zone at a time or times prior to the Late Devensian. Sutherland (1981a) argued that they may have been deposited during the Early Devensian during a period of ice-sheet expansion. The faunas contained in the silts and clays are high boreal to arctic in character (Horne *et al.,* 1894, 1897; Munthe, 1897; Jessen, 1905), implying the sediments were deposited during a glacial rather than an interglacial. At such a time, isostatic downwarping would be necessary for the sea to have access to the coastal zone. Amino acid analysis of mollusc fragments from various of the shell beds indicates that they were deposited in more than one event, however. Those at Clava and King Edward were deposited during the Middle or Early Devensian (Miller *et al.,* 1987; Merritt, 1990b), whereas those at Tangy Glen were deposited during the Middle Pleistocene (Gray, 1985; D. G. Sutherland, unpublished data), possibly between 250 ka and 300 ka. The fossiliferous clays at Burn of Benholm have amino acid ratios (Bowen, 1991) that suggest they were deposited, either as a till (Hall and Connell, 1991) or as a marine deposit (Sutherland, 1981a) during the Middle Pleistocene, possibly between 420 ka and 480 ka.

As with the Early Pleistocene marine event inferred from the presence of mollusc fragments in more recent glacial deposits in Aberdeenshire, so other marine events are implied by the age of shells contained in tills elsewhere around the Scottish coast. One such event appears to have occurred during the Middle Devensian. This has been suggested by amino acid ratios on mollusc fragments in till in Ayrshire (Jardine *et al.,* 1988) and amino acid ratios and radiocarbon dates on mollusc fragments from glacial deposits on Lewis (Sutherland and Walker, 1984).

A further group of marine deposits pre-dating the last ice-sheet is the non-fossiliferous cobble beaches reported from the Outer Hebrides and the Orkney Islands (see North-west Coast of Lewis and Muckle Head and Selwick) (Wilson *et al.,* 1935; Gailey, 1959; McCann, 1968; Peacock, 1984a; Sutherland and Walker, 1984; Selby, 1987). These occur at between 5 m and 15 m above present sea level and frequently rest upon eroded rock or, occasionally, till surfaces. The age of the deposits is not known, nor has it been established whether they represent one or more periods of marine deposition. By analogy with beach deposits in southern Britain, it has been suggested (von Weymarn, 1974) that the beaches are interglacial, but there is no direct evidence for this.

Whereas the marine depositional record is fragmentary in both time and space, marine erosional features, often spectacularly developed, can be traced for many kilometres around the coast of Scotland (Figure 2.5). Even more extensive marine erosion surfaces have been identified on the continental shelves to depths in excess of 100 m (Flinn, 1964, 1969; Sutherland, 1984c, 1987d; Stoker and Graham, 1985). On the coast, these erosional features typically occur as rock platforms and backing cliffs, the surfaces of which are often ice moulded or mantled with glacial deposits (Wright, 1911; Sissons, 1967a, 1982b; McCann, 1968). The glaciated platforms occur at altitudes ranging from the present intertidal zone to over 35 m OD and there is a pattern in their distribution that appears to relate to the centres of ice dispersal. Thus wide intertidal platforms or platforms only a few metres above present sea level are found around almost all of the Scottish coast (see North-west Coast of Lewis, Milton Ness, Dunbar, Kincraig Point and Northern Islay) except in areas of intense glacial erosion close to the former centres of ice dispersal, this pattern being apparent both in areas such as the Outer Hebrides and Skye as well as on mainland coasts. High rock platforms have a more restricted distribution, being most outstandingly developed around the Inner Hebridean islands and on parts of the neighbouring mainland coast (see Northern Islay). They also occur along parts of the east coast (see Dunbar). These high rock platforms do not occur in areas peripheral to the former ice centres, where glaciated platforms at between 5 m and 10 m are encountered (see North-west Coast of Lewis, Port Logan and Glenacardoch Point).

The times of formation of these erosional features remain speculative, and there is a very strong possibility that they have been occupied by the sea on more than one occasion, as have the platforms in the present intertidal zone. Sissons (1981a, 1982b) has suggested that the high platforms were formed when the Scottish ice-sheet was quite extensive, ice cover thus precluding platform formation towards the centres of ice dispersal. On this hypothesis, isostatic depression at the time of platform formation and subsequent rebound would explain the altitude of the features. As a consequence of isostatic tilting these shorelines would pass below present sea level in the areas peripheral to the centres of ice dispersal, thus explaining the apparent absence of the high platforms in these areas. The lower, glaciated platforms in the more peripheral areas may be of some antiquity, that in North-west Lewis, for example, being overlain by deposits of two distinct glacial phases as well as by the possible interglacial deposit of Toa Galson (McCann, 1968; Peacock, 1984a; Sutherland and Walker, 1984). Although many rock platforms and some marine deposits have been described as having been formed during interglacials, no unequivocal evidence is yet known from Scotland to support such an interpretation.

Glacial deposits

The early glacial record is only slightly better known than the marine record. As with the latter, the offshore Pleistocene deposits provide a fuller sequence of events and better chronological control than the onland deposits, but from neither area can the complete series of glaciations be inferred. The earliest evidence for glaciation has been found in the central North Sea Basin, where glaciomarine sediments occur between the top of the Jaramillo geomagnetic event (990 ka) and the Brunhes–Matuyama geomagnetic boundary (780 ka) (Sejrup *et al.,* 1987). These glaciomarine deposits are overlain by silts and clays containing a marine fauna indicative of interglacial conditions, above which the sediments indicate a return to a cold, glacial environment. Sejrup *et al.* (1987) correlated this glacial/interglacial/glacial sequence with Oxygen Isotope Stages 22/21/20, implying that the first glaciation occurred between 860 ka and 840 ka and the second cold period between 810 ka and 780 ka. It has not been established whether the glaciomarine sediments of the first glaciation were deposited from Scandinavian or Scottish ice (or both). If the correlation suggested earlier with the shallow-water marine sediments off the Scottish coast is correct, then this would favour deposition from Scandinavian ice.

The first appearance of glacial sediments in the North Sea Basin at this time is significant, for the oxygen isotope record, which is primarily an indicator of former ice volumes (Shackleton and Opdyke, 1973; Shackleton, 1987), indicates that there was a marked change in the magnitude and frequency of glacial events at approximately 750–800 ka (Ruddiman *et al,* 1986, 1989; Ruddiman and Raymo, 1988). Prior to this time, glaciations had been of shorter duration (with a dominant periodicity of 41,000 years) and of lesser magnitude. Subsequently, glacial cycles had a dominant periodicity of about 100,000 years and total ice volume at the glacial maxima was greater than previously. As will be discussed below, subsequent to this first glaciation the North Sea Basin was invaded by Scottish ice on at least five separate occasions. Correlation of the Early Pleistocene low sea-level events discussed above with periods of more limited glaciation on land (and by implication in the Highlands of Scotland) suggests at least three periods of such limited glaciation during the Early Pleistocene.

The earliest dated glacial deposits derived from Scotland also occur in the North Sea Basin, where they have been encountered in boreholes in the Forth Approaches. There, towards the top of the Aberdeen Ground Formation, a sedimentary sequence has been interpreted as having been deposited from a grounded ice-sheet in the west, with glaciomarine sediments occurring eastwards of the presumed maximum extent of the glaciation (Stoker and Bent, 1985). These deposits occur immediately above the Brunhes–Matuyama palaeomagnetic boundary, which suggests that the glaciation may have occurred during Oxygen Isotope Stage 18, between approximately 770 ka and 730 ka. Andrews *et al.* (1990) state that the till in the west-central North Sea Basin, which was assigned by Sejrup *et al.* (1987) to the Saalian glaciation, lies within the upper part of the Aberdeen Ground Formation and hence could be correlated with this earlier glacial event. No glacial deposits on land have been correlated with this glaciation.

The top of the Aberdeen Ground Formation is a major erosional surface which has been attributed to extensive glaciation of the North Sea Basin (Cameron *et* cll., 1987; Andrews *et al.,* 1990) by both Scottish and Scandinavian ice. Overlying this surface are interglacial marine deposits which are considered to correlate with the Holsteinian (Hoxnian), thus suggesting that the glaciation that produced the erosion surface was probably the Anglian, which may have occurred during Oxygen Isotope Stage 12, between 480 ka and 420 ka.

A number of glacial deposits onshore have been correlated with this glaciation, although none of these correlations rests on firm dating evidence. At Kirkhill in Buchan, basal fluvial sands and gravels contain erratics which were derived from the west and were probably transported by the same ice that deposited the basal till (Leys Till) in the nearby Leys quarry (A. M. Hall, this volume). These sediments are undated, but on the basis of likely minimum ages for the overlying glacial and interglacial deposits, they have been tentatively assigned to the Anglian glaciation (Connell and Hall, 1984a; Hall and Connell, 1991). The basal till at Bellscamphie (Hall, this volume) may also have been deposited at this time. Other evidence of early glaciation that may also correlate with the Anglian occurs at Fugla Ness in Shetland. If the interglacial deposits there are *in situ* and of Hoxnian age, as has been suggested by Birks and Ransom (1969), then the till described by Chapelhowe (1965) as underlying the interglacial sediments would be Anglian or older. At Dalcharn, 15 km east of Inverness, a lowermost till (Dearg Till) and overlying outwash gravels (Dearg Gravels) have been assigned to the Anglian (Merritt, 1990a), as the Dalcharn Biogenic Complex, which is partly developed in the Dearg Gravels, contains a pollen assemblage similar to that found at Fugla Ness.

In the North Sea Basin Quaternary succession, the Aberdeen Ground Formation is overlain by the Ling Bank Formation and then, unconformably, by the Fisher Formation. Within these last two formations there is evidence of two glacial phases separated by a marine event (Andrews *et al.,* 1990). During both these phases, ice in the west-central North Sea Basin originated in the Scottish Highlands. The glacial and glaciomarine sediments deposited during the glaciations are underlain and overlain by interglacial deposits. The lower interglacial has been tentatively correlated with the Holsteinian (Hoxnian) and the upper interglacial correlated with the Eemian (Ipswichian), and hence the glacial sediments are considered to have been deposited during two phases of the Saalian ('Wolstonian'), between 180 ka and 130 ka. However, the occurrence of two quite distinct phases to the Saalian glaciation has also been noted in northern Europe (Ehlers *et al.,* 1991), where it has been suggested that they may represent two quite distinct glaciations, the former in Oxygen Isotope Stage 8, between 300 ka and 245 ka, and the latter in Oxygen Isotope Stage 6, between 180 ka and 130 ka. Resolution of the number and age of glaciations in this time interval awaits further work.

A number of glacial deposits have been assigned to this period between the Hoxnian and Ipswichian. At Kirkhill, a till, which is underlain by deposits containing a soil horizon tentatively correlated with the Hoxnian (Hall, 1984b; Hall and Connell, 1991), has undergone interglacial weathering, presumably during the Ipswichian. This till may be correlated with other weathered tills in north-east Scotland, possibly at Boyne Bay (Peacock, 1966) and with the middle till at Bellscamphie (Hall, this volume). In the Inverness area, the Dalcharn Lower Till overlying the Dalcharn Biogenic Complex, the Suidheig Till underlying the Odhar Peat, and the Cassie Till at the base of the Clava succession have all been tentatively assigned to this period (Merritt, 1990a). More widely, the event represented by these deposits may correlate with the early phase of glaciation on the North-west Coast of Lewis which pre-dates the possible interglacial site of Toa Galson (Sutherland and Walker, 1984) and which may have been responsible for the early transport of erratics on to the outer islands such as St Kilda (Sutherland, 1984a; Sutherland *et al.,* 1984). The glacial deposits on the western shelf edge identified by Stoker (1988) together with the earlier moraine system described by Stoker and Holmes (1991) may also date from this period, as may the basal till at Leavad in Caithness.

These glacial deposits all pre-date at least one interglacial or are overlain by sediments attributed to the Late Devensian glaciation. It has been suggested that rather than only one phase of ice-sheet glaciation post-dating the Ipswichian (i.e. during the Late Devensian) there was a build-up of an ice-sheet in Scotland during the Early Devensian (Sutherland, 1981a). There is as yet no unequivocal evidence for glaciation in Scotland during the Early Devensian (Bowen *et al.,* 1986), although a number of glacial deposits have been tentatively attributed to this period. These include the uppermost till at Kirkhill (Hall, 1984b; Hall and Connell, 1991), the Moy Till and the lower member of the Dalcharn Upper Till in the Inverness area (Merritt, 1990a) and the basal till at Sourlie in Ayrshire (Jardine *et al.,* 1988). Results of amino acid analysis of shells contained in the till in Caithness and Orkney have been interpreted as supporting the hypothesis of Early Devensian glaciation (Bowen and Sykes, 1988; Bowen, 1989, 1991), but no stratigraphic evidence has been found in these areas for such an event (cf. Hall and Whittington, 1989; Hall and Bent, 1990).

Periglacial deposits

Pre-Late Devensian periglacial deposits have not been frequently described, although it seems very likely that in the lowlands, in particular, cold non-glacial periglacial climatic regimes occurred in Scotland during much of the Pleistocene. The earliest such deposits known are those at the base of the Kirkhill sequence; and periglacial deposits also occur overlying the lower palaeosol at this site, formed prior to deposition of the weathered till. Periglacial reworking has also affected the Dalcharn Biogenic Complex sediments, which contain a pollen assemblage that may correlate with the Hoxnian. Above the weathered till at Kirkhill, but under the uppermost glacial deposits, there are further periglacial sediments (Connell and Hall, 1984a, 1987). These may have formed during the Early Devensian, if the attribution of the weathering of the underlying till to the Ipswichian is correct. Other putative Early Devensian periglacial deposits are those immediately overlying the possible Ipswichian palaeosol at Teindland, those at the base of the Castle Hill sequence at Gardenstown (Sutherland, 1984b), the sediments of the Moy Paraglacial Complex overlying the Allt Odhar interstadial peat (Merritt, 1990c) and those overlying the possible Ipswichian peat at Toa Galson on the North-west Coast of Lewis (Sutherland and Walker, 1984). Pre-Late Devensian soliflucted deposits are also present at Sel Ayre and Fugla Ness on Shetland, where they underlie inferred Late Devensian tills (Mykura and Phemister, 1976; A. M. Hall and J E. Gordon, unpublished data).

Non-marine organic sediments

Non-marine organic sediments that pre-date the Late Devensian have only been located at a small number of localities in Scotland (Figure 2.6). They comprise peats, lacustrine sediments and palaeosols, and a small number have been found in association with mammal remains. They are rarely of any great thickness and almost all have been truncated by erosion. Some are considered to have been deposited during interglacials, others during interstadials.

The best preserved interglacial deposit is possibly also the oldest, a peat at Fugla Ness in Shetland. This has a clear thermophilous vegetational succession and macrofossils of trees have been recovered from the peat. It has been correlated with the Hoxnian (Birks and Ransom, 1969), but such an attribution must remain uncertain (Lowe, 1984), not least because of the considerable distance of Shetland from the reference sites for this interglacial in south-east England and the extent of vegetational differences between these areas today. The vegetational record from Fugla Ness shows a strong oceanic influence and certain species are strongly suggestive of a Middle Pleistocene age. Comparison with the oceanic climatic record for the North Atlantic reveals only two periods in the last 600 ka that were as warm or warmer than the Holocene, the Ipswichian and a period at around 380 ka (Oxygen Isotope Stage 11) (Ruddiman and McIntyre, 1976). Thus the Fugla Ness peat may have been formed during that earlier period (Sutherland, 1984a).

At Dalcharn, near Inverness, organic sediments comprising the Dalcharn Biogenic Complex (Merritt and Auton, 1990) have been shown to contain a pollen assemblage rather similar to that found at Fugla Ness. The presence of significant quantities of *Ilex* (holly) pollen, in particular, implies an interglacial origin, and its abundance at Dalcharn, which is close to the present northern limit of *Ilex,* may suggest a climate somewhat warmer than today (Walker, 1990a). On these grounds, tentative correlation may be made with the Hoxnian.

The lower palaeosol at Kirkhill and the immediately overlying organic sediments were originally taken to have been formed during an interglacial (Connell *et al.,* 1982), but more recent work has shown that the palaeosol is most akin to a cold-water gley which could develop today in Scotland above 900 m OD (Connell and Romans, 1984). Hence, at Kirkhill (altitude, 50 m OD), the palaeosol probably formed during an interstadial. However, detrital organic material contained in the overlying sands was eroded from a land surface with a temperate soil and vegetation cover (Connell, 1984a; Lowe, 1984) and thus, although no *in situ* interglacial deposits are to be observed at Kirkhill, an interglacial period can be inferred to have occurred prior to the development of the palaeosol. From its stratigraphic position, this interglacial has been correlated with the Hoxnian.

Other interglacial (or presumed interglacial) deposits have proved difficult to correlate. The principal reason is that the pollen spectra at these sites have generally revealed contemporaneous vegetational communities which were dominated by grasses and sedges, with acid heath development at certain sites. Tree pollen, which has been used to recognize and differentiate interglacial deposits in more southerly areas of Britain, is present in these pollen spectra in only very low amounts and can be attributed to long-distance transport. Thus the organic sediments at Sel Ayre (Birks and Peglar, 1979), Teindland (FitzPatrick, 1965; Edwards *et al.,* 1976) and Toa Galson (North-west Coast of Lewis) (Sutherland and Walker, 1984), each of which may have been deposited, at least in part, during the Ipswichian, cannot be attributed to that interglacial with any confidence (Lowe, 1984). The truncated palaeosol developed in the lower, weathered till at Kirkhill has also been assigned to the Ipswichian.

Pre-Late Devensian interstadial deposits, though still rare, are relatively more abundant than presumed interglacial deposits. With the exception of the interstadial represented by the lower palaeosol at Kirkhill discussed above, all the interstadial deposits have been attributed to the Devensian. They may be divided into two groups: those that are beyond the range of radiocarbon dating and which are likely to be Early or early Middle Devensian, and those that yield apparently finite radiocarbon dates from the Middle Devensian.

The most detailed record of a presumed Early Devensian interstadial site is that from the Odhar Peat at Allt Odhar in Inverness-shire. Pollen analysis of the peat (Walker, 1990b) has demonstrated a vegetational succession from birch woodlands with willow and juniper scrub interspersed with open grassland, to heathland and, finally, to an open vegetation dominated by species-poor grass and sedge communities. Pollen of taxa of northern or montane affinities are present throughout the deposit. Radiocarbon dating has indicated that the peat was formed before 51,000 BP and probably earlier than 62,000 BP (Harkness, 1990). An initial uranium-series date on the peat gave an age of 124 ± 13 ka (Heijnis, 1990) but the revised age estimate is 106 + 11/–10 ka, which places the peat in an Early Devensian interstadial, equivalent to Oxygen Isotope Substage 5c (Walker *et al.,* 1992).

An Early Devensian interstadial origin has been proposed for the peat lens in till at Burn of Benholm. This has a pollen spectrum of low floristic diversity dominated by grasses and sedges. It has a radiocarbon age of >42,000 BP (Donner, 1960, 1979). The limited information on this deposit precludes attempts at wider correlation.

Several sites with radiometric ages between approximately 26,000 BP and 35,000 BP appear to relate to an interstadial in the later part of the Middle Devensian. Various mammal remains (reindeer, woolly rhinoceros) have also been dated from this period. The most detailed record to have been published to date is that from Tolsta Head in Lewis (von Weymarn and Edwards, 1973; Birnie, 1983), for which a relatively cool climate and unstable soil conditions are suggested. Deposits at Sourlie (Jardine *et al.,* 1988) also date from this period, as do reindeer bones recovered from the caves at Creag nan Uamh (Lawson, 1984) and the woolly rhinoceros bone found in glacial sediments at Bishopbriggs (Rolfe, 1966). Although none of them have been radiocarbon dated, it is likely that at least some of the mammoth remains found in central Scotland are also of this age. Further Middle Devensian interstadial sites are those at Crossbrae Farm in north-east Scotland (Hall, 1984b) and possibly Abhainn Ruaival on St Kilda (Sutherland *et al.,* 1984). Interstadial pollen spectra from Teindland have also been attributed to this phase (Edwards *et al.,* 1976), but this has been disputed (Sissons, 1981b; Lowe, 1984) and an alternative explanation of an Early Devensian age advanced (Sissons, 1982c). During this interstadial, speleothem deposition occurred in caves in Sutherland (Atkinson *et al.,* 1986) and, as noted earlier, there was marine invasion of the Scottish coast.

In summary, the pre-Late Devensian Pleistocene record is only partially known and its fragmentary nature in the absence of radiometric dates makes correlation between sites a matter of some speculation. The available evidence can be synthesized into a scheme involving three and possibly five periods of ice-sheet glaciation (Figure 2.7) and certain other periods of limited glaciation during the Early Pleistocene may also be inferred. Marine sedimentation along the inner coasts has taken place on two (and probably more) occasions, though the greater part of the marine record, consisting of erosion platforms and non-fossiliferous beaches, remains as yet undated. The non-marine interglacial sites can be interpreted as representing two separate interglacial stages, although these correlations are tentative. An interstadial succeeding the earlier of those interglacials is known from 1Kirkhill, and other interstadial sites have been assigned to the Devensian. At least one Early Devensian inter-stadial deposit has been proposed and deposits at a number of sites throughout the country appear to be from an interstadial towards the end of the Middle Devensian. No sites have been found that have been correlated with the Upton Warren interstadial in England. Lowland periglacial deposits have been identified from at least three periods. By comparison with the sequence of Pleistocene climatic change derived from both deep-sea cores (Ruddiman and McIntyre, 1976; Zimmerman *et al.,* 1984) and the terrestrial record in Holland (Zagwijn, 1975, 1986) and elsewhere in Britain (Bowen *et al.,* 1986; Ehlers *et al.,* 1991), it is apparent that there are very many events not represented in the present synthesis of the Scottish evidence. This implies that much remains to be discovered and, further, that the above synthesis is probably an over simplification. It is therefore clear that in the apparent absence of so much evidence the scientific value of the known sites is at a premium.

Late Devensian ice-sheet glaciation

Ice-sheet chronology, flow patterns and dimensions

The locations of the sites at which late Middle Devensian interstadial deposits have been found indicate that much of lowland and coastal Scotland was ice-free during the period between about 35,000 BP and 26,000 BP. The climate, however, appears to have been more severe than at present and hence there is a strong possibility that glaciers would have existed in the Highlands at this time. Subsequently, there was a major expansion of ice, initially from the main ice centre of the western Highlands and the neighbouring islands such as Skye and Mull and later, as the glaciation progressed, from the ice centre in the Southern Uplands (Figure 2.8).

The chronology of this period of glaciation is not known in detail, but a radiocarbon date on ice-transported mollusc fragments, if correct, suggests that the independent Outer Hebridean ice-cap only expanded beyond the present coast after 23,000 BP (Sutherland and Walker, 1984). In the North Sea Basin, the Marr Bank Formation was deposited in a shallow-water glaciomarine environment east of the ice-sheet margin (Sutherland, 1984a; Stoker *et al.,* 1985) and a radiocarbon date of 17,730 ± 480 BP (SRR–625) indicates that the ice was close to its maximum at this time. Glaciomarine sediments at St Fergus in Aberdeenshire, deposited during ice retreat, have been radiocarbon dated to 15,320 ± 200 BP (Lu–3028) (Hall and Jarvis, 1989). Farther south, five raised shorelines between Stonehaven and Fife Ness (Cullingford and Smith, 1980) which were formed in succession during ice retreat (see Dryleys, and Milton Ness) can be dated to between 16,000 BP and 14,000 BP (Sutherland, 1991a) on the basis of a calculation of the change in shoreline gradient with shoreline age (Andrews and Dugdale, 1970). Radiocarbon dating of basal sediments in deglaciated areas (see Cam Loch, Loch Etteridge, and Loch an t-Suidhe), although potentially subject to error (Sutherland, 1980), together with similar dating of marine molluscs in deposits laid down upon retreat of the ice-sheet (see Geilston), implies that by around 13,000 BP almost all the ice had retreated to within the Highland boundary. Thus the whole period of lowland glaciation, both advance and retreat, may have lasted only around 10,000–12,000 years.

Not all of Scotland was covered by ice during this period as ice-free areas have been identified in both North-west Lewis (Sutherland and Walker, 1984) and around Crossbrae Farm near Turriff (Hall, 1984b), although the precise limits of the latter ice-free area are unclear (see Castle Hill and Boyne Quarry) (Sutherland, 1984a; Hall, 1984b; Hall and Bent, 1990). It has also been suggested that much of Caithness (see Baile an t-Stratha and Drumhollistan) and Orkney (see Den Wick and Mill Bay) were not glaciated at this time (Sutherland, 1984a), but there is no direct evidence for this, Hall and Whittington (1989) and Hall and Bent (1990) arguing for complete glaciation during the Late Devensian.

At the maximum of the glaciation two major lobes of ice appear to have extended into the North Sea Basin, one emerging from the Midland Valley and the other from the inner Moray Firth. Both terminated in end-moraine complexes fronting into a shallow arctic sea (Sutherland, 1984a; Cameron *et* al., 1987; Hall and Bent, 1990), with a large part of the central North Sea being dry land. Off the west coast, the ice-sheet terminated to the north of Lewis (Sutherland, 1991a), with a larger lobe extending towards the shelf edge to the south of the Hebridean islands (Selby, 1989; Peacock *et al.,* 1992). Low sea levels on the outer western continental shelf apparently also imply extensive areas of dry land at this time (Sutherland *et al.,* 1984; Sutherland, 1984c, 1987d). By 15,250 BP, the St Kilda Basin was free of glacier ice (Peacock *et al.,* 1992). During the Late Devensian, Shetland appears to have been covered by an independent ice-cap (Sutherland, 1991b). Comparison of models of glacio-isostatic rebound (Lambeck, 1991a, 1991b), with observed patterns of raised shorelines, also suggests that the ice in Scotland was less extensive about 18,000 years ago than suggested in some ice-sheet models (Boulton *et al.,* 1977, 1985; Denton and Hughes, 1981).

The presence of ice-free areas along parts of the northern fringes of the ice-sheet, its relatively modest dimensions (the flow-line distance from ice shed to ice margin in the North Sea Basin was about 200 km) and the coexistence of ice-caps nurtured on relatively small mountain masses all imply that the main Highland ice mass was not particularly thick at its maximum (see also Lambeck, 1991a, 1991b). Striations and erratics have been reported at up to 1100 m OD in the Ben Nevis range and, although they cannot be dated, they may be attributed to the last ice-sheet, suggesting that it overtopped the mountains in that area (Sissons, 1967a; Thorp, 1987). Similarly, erratics found on the top of Merrick, *in* the western Southern Uplands, suggest that the ice there also overtopped the highest hills (Charlesworth, 1926a; Cornish, 1982).

In the Northern Highlands, however, it is probable that the highest mountains stood above the ice-sheet as nunataks (see An Teallach and Ben Wyvis) (Godard, 1965; Ballantyne, 1984; Sutherland, 1984a; Ballantyne *et al.,* 1987; Reed, 1988), and at that time some of the major periglacial features on these mountain summits may have been formed. Ice-surface altitudes of around 700 m OD on An Teallach rising to over 800 m OD in the Fannich mountains have been suggested (Ballantyne *et al.,* 1987). On the Trotternish Peninsula on Skye, Ballantyne (1990) has mapped a periglacial trimline considered to represent the upper limit of the Late Devensian ice-sheet. This trimline descends from approximately 600 m to around 450 m OD in 24 km from south to north along the peninsula. In northern Lewis, the moraine that Sutherland and Walker (1984) identified as marking the edge of the last ice-sheet descends from over 110 m to around 50 m OD in 6 km.

Within the limits of the ice-sheet, till was deposited, this being almost continuous in the lowland areas, where it frequently forms drumlins. In different parts of the country the till contains specific suites of erratics, largely derived from the immediate locality but also containing material from much further afield. The distribution of these erratics (see Clochodrick Stone), together with the orientation of ice-streamlined landforms and striations (see Agassiz Rock), gives a graphic picture of the ice-flow patterns (Figure 2.8) and allows the relative strengths of the ice flow from different ice centres to be assessed.

The major ice mass was centred in the western Highlands, the ice shed extending in an approximately north–south alignment from Sutherland to the Cowal Peninsula in Argyll (Sissons, 1976b; Boulton *et al.,* 1991). Ice flowed outwards from this area, transporting erratics of Highland aspect throughout almost the whole of the Central Lowlands and even into the northern Southern Uplands. As the glaciation progressed, however, the Southern Uplands ice expanded and exerted sufficient pressure to force back the Highland ice throughout the southern Central Lowlands. This resulted in the classic two-fold stratigraphy in this 'debatable ground' of a till containing Southern Upland erratics overlying a till containing Highland erratics (see Nith Bridge, Hewan Bank and Port Logan). The Highland ice also abutted against ice flowing away from the independent ice domes that developed on Skye (see the Cuillin), Mull, the Outer Hebrides, the Cairngorms and the south-east Grampians but did not overwhelm any of them.

Ice-sheet deglaciation

The retreat of the last ice-sheet was accompanied by the release of abundant meltwater, which reworked debris both beneath and held within the ice. Intricate sequences of meltwater channels were often cut across spurs and other areas of positive relief, these being recognizable by their independence from the present drainage system, their anastomosing pattern in plan, their frequent up-and-down long profiles (resulting from formation under hydrostatic pressure below the ice) and, at times, their very large size (see Rammer Cleugh, Carlops, Corrieshalloch Gorge, Glen Valtos, the Cairngorms and Muir of Dinnet).

In topographic depressions and valley bottoms, the depositional counterparts of the meltwater channels are extensive sequences of bedded sands and gravels, often with very characteristic morphologies. Where deposited by rivers flowing through channels below or within the ice they possess an elongate, sinuous morphology, and frequently groups of these eskers form complex anastomosing patterns. The most noted of such features occur at Carstairs, but there are also outstanding examples at Kildrummie, Torvean, Littlemill and Kippet Hills. Other areas of bedded sands and gravels have a locally more complex morphology with frequent rounded, steep-sided hollows where individual masses of 'dead' ice became surrounded by sand and gravel, the hollows (kettle holes) being produced when the 'dead' ice melted (see Muir of Dinnet and the Cairngorms). The intervening mounds of sand and gravel are frequently flat-topped and, where banked against the sides of a valley, have a clear terrace form. These flat-topped kames and kame terraces can often be traced into outwash sands and gravels deposited in front of the decaying ice, this relationship indicating the role of meltwater drainage through or around the 'dead'-ice masses in forming these deposits. Outstanding examples illustrating these relationships are found at Torvean and Moss of Achnacree (although the latter example was formed during retreat of a Loch Lomond Readvance glacier, see below), and an excellent series of outwash terraces occurs along the North Esk near Edzell.

The various meltwater landforms and deposits did not form randomly in the landscape and when regional patterns of meltwater channels and bedded sands and gravels are mapped they reveal the major drainage routes across the country of the decaying ice-sheet (Figure 2.9) (Sutherland, 1984a, 1991a). Thus meltwater drainage from the hills of south-central Ayrshire was eastwards across the central Clyde valley and, via the Midlothian Basin, into the North Sea. Similarly, meltwaters from the Tay Valley initially drained north-eastwards through Strathmore to reach the coast near

Montrose.

Corresponding to the coarse-grained sand and gravel deposits found inland are fine sands, silts and clays deposited along the coasts, most especially in the eastern estuaries and in the offshore zone during this period of ice-sheet retreat (Figure 2.10). Along much of the east coast these fine-grained deposits have long been famous for the high-arctic character of the marine molluscs, ostracods and Foraminifera contained within them (Jamieson, 1865; Brown, 1868; Brady *et al.,* 1874; Davidson, 1932; Graham and Gregory, 1981; Paterson *et al.,* 1981). They have been studied in most detail in the Inchcoonans and Gallowflat area beside the Tay estuary, where the Errol Clay Pit has given the sediments their informal name of the Errol beds (Peacock, 1975c). The offshore equivalents in the North Sea are termed the St Abbs Formation in the Forth Approaches (Stoker *et al.,* 1985) and the Fladen Member of the Witch Ground Formation in the central North Sea Basin (Long *et al.,* 1986). Radiocarbon dating of the North Sea sediments, together with the close association of the Errol beds with the sequence of raised shorelines formed as the ice retreated, indicates that the Errol beds were deposited from some time prior to 16,000 BP until approximately 13,000 BP. Deposition of sediments containing high-arctic faunas during a large part of the retreat of the last ice-sheet implies that the climate was still very cold and therefore that the retreat of the ice was not a consequence of warmer temperatures. It has therefore been inferred that a reduction in precipitation was responsible for the initial period of ice decay (Sutherland, 1984a).

The estuarine sediments generally coarsen towards the rivers that supplied the sediments and at these river mouths complex sedimentary sequences often developed. Sea-level fall in response to isostatic uplift was a major control on these sequences, with coarse-grained fluvial or outwash sediments interdigitating with or being deposited on top of the estuarine sediments as the shoreline retreated. In the Perth area, for example, such a sequence of outwash gravels overlying estuarine silts was originally thought to represent a readvance of the ice-sheet (the so-called Perth Readvance — Simpson, 1933; Sissons, 1963a). Paterson (1974), however, has demonstrated that the sequence need not represent a readvance but simply results from the deposition of outwash sediments across the estuarine silts as sea level fell (see Almondbank).

One of the major consequences of ice-sheet glaciation was glacio-isostatic depression of the Earth's crust, such depresssion being greatest where the ice was thickest in the western Highlands. This depression of much of Scotland was significantly greater than the glacio-eustatic lowering of world sea level due to the contemporaneous expansion of the North American and Scandinavian ice-sheets, so that the rebound upon removal of the ice load has resulted in a sequence of raised shorelines around much of the coasts of mainland Scotland and the Inner Hebrides (Sissons, 1967a; Gray, 1985). It is only around the coasts of the outer islands that the eustatic rise of sea level has been greater than any isostatic effect and where, as a consequence, shorelines that date from the deglacial phase are now submerged.

These raised shorelines frequently form prominent elements of the coastal scenery, none more so than the 'staircases' of white, unvegetated, quartzite shingle ridges along the West Coast of Jura. The finer-grained sediments in the low-energy estuaries of the east coast formed extensive sequences of low-gradient terraces, which are well displayed at Munlochy Valley, Dryleys and Inchcoonans and Gallowflat. The progressive decline in the rate of isostatic uplift since deglaciation has meant that terraced sequences of distinct shorelines formed, each younger shoreline occurring in a given area at a lower altitude and with a lower gradient than its predecessor. Particularly along the east coast where individual shorelines have been mapped, the younger features are found to extend progressively farther west, reflecting a period of continous retreat of the ice-sheet (Cullingford and Smith, 1966, 1980; Sissons *et al.,* 1966; Firth, 1989a), during which much of the Central Lowlands were deglaciated.

Although various halts or even readvances have been postulated during the period of ice-sheet retreat, almost all of these have now been discounted. An important exception occurs in the Northern Highlands where a major end moraine (see Gairloch Moraine) has been identified as marking the limit of a readvance that interrupted the later part of ice retreat (Robinson and Ballantyne, 1979; Sissons and Dawson, 1981; Ballantyne *et al.,* 1987). This readvance has been associated with a distinct shoreline which is partly cut in rock and it is possible that the readvance dates from the period around 13,500–13,000 BP when the climate changed from arctic cold to a more boreal regime at the opening of the Lateglacial Interstadial. Another halt in the ice retreat on the west coast may be marked by an end moraine complex identified on Islay (see Northern Islay) (Dawson, 1982) which Sutherland (1991a) has suggested formed at around 14,500 BP. Near the mouths of numerous sea lochs along the west coast there are marked drops in the marine limit

(Peacock, 1970a; Sutherland, 1981b) which could be the result of glacier readvance but are equally likely to reflect topographic control on the rate of calving of the glaciers flowing down the sea lochs. Minor local readvances of ice may also be represented in the Perth area at the Shochie Burn site and near Inverness at Ardersier.

Errol beds sediments with their high-arctic faunas are absent along much of the west coast of Scotland, only being recorded in the North Minch (Graham *et al.,* 1990) and, possibly, around Stranraer (Brady *et al.,* 1874). Immediately upon deglaciation along the west coast, marine sands, silts and clays containing boreo-arctic faunas (the Clyde beds; see Geilston) were deposited. The absence of Errol beds has led to the suggestion that these areas were ice covered until the climate ameliorated (Sissons, 1967a; Peacock, 1975c). A corollary to this hypothesis is that a large ice dome existed across the Firth of Clyde until a very late stage of deglaciation, which may explain the easterly and westerly to northwesterly flow of ice out of this area at that time. Holden (1977a) has postulated a late readvance of ice from the Clyde area into central Ayrshire to explain a tripartite sedimentary sequence near Greenock Mains, and late-stage flow from the south-west across Jura is indicated by the Scriob na Caillich medial moraine on the west coast of that island (Dawson, 1979b). It has been suggested (Sutherland, 1984a) that rapid deglaciation of the Firth of Clyde occurred slightly before 13,000 BP, as radiocarbon dates on marine shells from the head of the firth and the Clyde estuary imply deglaciation by that date (Sutherland, 1986).

Lateglacial Interstadial

The Lateglacial Interstadial was a period of mild climate which opened around 13,000 BP. As the oceanic polar front in the North Atlantic migrated to the north of Scotland (Ruddiman and McIntyre, 1973, 1981a; Ruddiman *et al.,* 1977), North Atlantic Drift waters reached the Scottish coast (Peacock and Harkness, 1990) and both marine and atmospheric temperatures rose dramatically to close to present-day values (Bishop and Coope, 1977; Peacock, 1981b, 1983a, 1989b; Atkinson *et al.,* 1987). The change in the nearshore marine environment from high arctic to boreal may have occurred in only 50 years (Peacock and Harkness, 1990).

This rapid change in climate was not immediately registered in the nature of the vegetation because of the time necessary for soil development and plant expansion. Plant communities, as indicated by pollen records at sites throughout Scotland (such as Loch an t-Suidhe, Loch of Winless, Cam Loch, Din Moss, Loch Etteridge, Burn of Aith and Loch Cill an Aonghais) (Figure 2.11), were characterized by open-habitat taxa typical of pioneer vegetation on recently deglaci-ated terrain. A much more rapid response to climatic change has been revealed both by Coleoptera in the terrestrial environment and by marine faunas. It is only in southern Scotland that beetle assemblages from the Lateglacial Inter-stadial have been studied (see Redkirk Point and Bigholm Burn) and they have revealed that around 13,000 BP summer temperatures were close to present-day values (Bishop and Coope, 1977). Analysis of marine faunal assemblages has also suggested that water temperatures at around 12,800 BP were within 1–2°C of those of the present (Peacock, 1981b, 1983a; Peacock and Harkness, 1990).

As the interstadial progressed, plant succession led to the development of a closed vegetation cover throughout the lowlands and in the Highland valleys (Walker, 1984b; Tipping, 1991a). There was considerable geographical diversity in the interstadial plant communities, reflecting the varied topographic, edaphic and climatic controls on plant growth. At a number of sites, however, there was an apparent brief halt in the progressive establishment of the vegetation, shown by an increase in the representation in the pollen spectra of open-habitat taxa, a reduction in woody plants and, in places, an increase in the inwashing of mineral matter (see Stormont Loch, Pulpit Hill, Cam Loch, Loch an t-Suidhe and Loch Sionascaig). This phase, which has not been observed at all sites investigated (Tipping, 1991a, 1991b), has been correlated with the Older Dryas chronozone (12,000–11,800 BP) of the north-west European sequence (Pennington, 1975b). However, accurate radiocarbon dating at a number of sites to establish the age of the event and whether it does indeed correlate with the Older Dryas has not been carried out. On the assumption that it is a single event, Walker and Lowe (1990) have suggested that its registration at some sites but not at others may be a reflection of changes in seasonal temperatures, as Atkinson *et al.* (1987) reported that at about this time in the interstadial there were periods when winter temperatures decreased although summer temperatures remained broadly unchanged.

Subsequently, there developed throughout the country a complex mosaic of vegetational communities during the main part of the interstadial. In south-east and eastern Scotland open birch woods with subordinate juniper occurred (see Beanrig Moss, Stormont Loch and Muir of Dinnet) and along the western seaboard as far north as Mull (Bigholm Burn, Loch Cill an Aonghais, Loch an t-Suidhe) grassland communities were dominant, with local occurrence of juniper and *Empeum* (crowberry) heaths and tree birch. In the north and west (Loch Sionascaig, Cam Loch) dwarf-shrub heaths were dominant, interspersed with grasslands, whilst in the central Highlands (Loch Etteridge, Abernethy Forest) a shrub tundra with *Empetrum, Betula* (birch) and some *Salix* (willow) occurred. In the valleys of the south-east Grampians a closed grassland developed with juniper, willow, dwarf birch and, in sheltered areas, tree birch. Moss heaths and floristically poor grasslands were present on upper slopes.

Tree development during this part of the interstadial was limited to birch and pine, which occurred only in particularly favoured localities (Gray and Lowe, 1977b). Tree birch had a northern limit on the west coast in southern Skye (Loch Ashik, Loch Meodal) but it was also present along the east coast into Aberdeenshire and possibly even Caithness (Loch of Winless). Pine appeared only locally in the eastern central lowlands and Aberdeenshire (Black Loch, Muir of Dinnet).

The broad pattern of vegetation that developed during the interstadial shows vegetational zones which have a geographical expression similar to the vegetation zones that developed under the different climatic conditions of the middle Holocene (Birks, 1977; Bennett, 1989) and which have been identified in the relict vegetation of today (McVean and Ratcliffe, 1962).

The vegetational succession would appear to indicate that, for the greater part of the interstadial, the climate became milder (with a possible brief 'revertance' phase). As with the opening of the interstadial, however, environmental indicators with a faster response time to climatic change than plants suggest otherwise and in both the coleopteran and marine faunal records the main part of the interstadial has been found to be significantly cooler (by 2–3°C) than its opening phase shortly after 13,000 BP. It may be speculated that since such a drop of temperature would be compatible with glaciation of the highest mountains (Manley, 1949), then glaciers would have been likely to exist in the Highlands through the Lateglacial Interstadial. Sutherland (1987c) has suggested that the laminated sediments deposited throughout the interstadial at Loch Droma in Wester Ross (Kirk and Godwin, 1963) may be evidence for the continued existence of glaciers in the adjacent mountains.

Towards the end of the interstadial (between approximately 11,500 BP and 11,000 BP) both the vegetational and the coleopteran records indicate that a decline in temperature was under way as a precursor of the Loch Lomond Stadial. Interestingly, the nearshore marine faunal record indicates that there was a brief period at about 11,250–11,000 BP of warmer sea-surface temperatures (Peacock, 1981 b, 1987, 1989b; Peacock and Harkness, 1990). It has been suggested (Peacock, 1987, 1989b) that this marine warming was due to a brief strengthening of the North Atlantic Drift as the oceanic polar front began to move southward in the later part of the interstadial. It can be speculated that this conjunction of a relatively warm sea and a cooling land (cf. Ruddiman and McIntyre, 1979; Ruddiman *et al.,* 1980) would have been particularly favourable for glacier development (Sutherland, 1984a).

During the early part of the Lateglacial Interstadial, sea level around most of the Scottish mainland coast, and particularly around the Highland margins, was falling rapidly. In the more peripheral areas, sea level was below its present level throughout this period and freshwater deposits accumulated within (and presumably below) the present intertidal zone (see Redkirk Point). In the later part of the interstadial, sea level everywhere around the Scottish coast was either below its present level or below the level attained following the Main Postglacial Transgression (see below), so that the course of sea-level change can only be deduced indirectly. Thus, from inferences based on the depth of water in which marine micro- and macrofaunas would have lived, Peacock *et al.* (1977, 1978) found that in the area of the sea lochs at the head of the Firth of Clyde, sea level was relatively stable during this period.

In summary, the Lateglacial Interstadial is one of the most intensively studied periods during the Scottish Late Pleistocene and a wealth of data have been accumulated on the natural environment and its evolution during the 2000 years after climatic amelioration at around 13,000 BP. The prime driving force of environmental change has been established to be the movement of the oceanic polar front in the North Atlantic. The studies that have been carried out have been rewarded with insights into the response of different parts of the environment to a common, major climatic stimulus followed by a period of mild but cooler climate than at present.

The Loch Lomond Stadial

At around 11,000 BP, the climatic decline that had become apparent towards the end of the Lateglacial Interstadial intensified and there was a return to arctic conditions during the Loch Lomond Stadial. This period of severe climate, which lasted approximately 1000 years, has been registered by all palaeoenvironmental indicators that have been studied. It corresponds with a period of glacier readvance, a return to tundra and open-habitat plant communities throughout the country, the destruction of the soil profiles that had developed during the preceding inter-stadial and accompanying inwashing and solifluction of fine sands, silts and clays into lakes and enclosed basins, a return of polar waters to the neighbouring seas and the formation or reactivation of large-scale periglacial landforms and deposits (Sissons, 1979e).

Glaciation

The precise chronology of the glacier readvance (Loch Lomond Readvance) that reached its maximum during the climatic deterioration at the end of the Lateglacial has not been established. It is likely that glaciers began to readvance during the decline in climate at the end of the Lateglacial Interstadial. Radiocarbon dates on marine molluscs and terrestrial organic sediments that were transported or overridden by outlet glaciers of the western Highland ice-cap (see Western Forth Valley, Gartness, Croftamie, Rhu Point and South Shian and Balure of Shian) as well as on the Isle of Mull have confirmed the general age of the readvance to be between 11,000 and 10,000 BP (Sissons, 1967b; Peacock, 1971c; Gray and Brooks, 1972; Rose, 1980c, 1980e; Browne and Graham, 1981; Sutherland, 1981b; Browne *et al.,* 1983; Rose *et aL,* 1988; Peacock *et al.,* 1989; Merritt *et al.,* 1990). A similar inference that the readvance maximum occurred during the Loch Lomond Stadial can be made from a comparison of sediment sequences in enclosed basins 'inside' and 'outside' the readvance limit (Donner, 1957; Sissons *et al.,* 1973), the former typically containing only sediments from the end of the stadial and the Holocene, whereas the latter frequently have complete Lateglacial profiles in addition to Holocene sediments (see Mollands, Tynaspirit, Pulpit Hill and Loch Etteridge).

Details of the chronology of ice advance and retreat and the date of the maximum extent of the readvance are not well known, and locally, at least, it is possible that the glaciers did not all reach their maximum positions at the same time (Bennett, 1990). At Croftamie, Rose *et al*. (1988) have dated organic sediments underlying deposits of the Loch Lomond glacier, which have indicated that the readvance culminated after 10,500 BP in this area. This chronology is in agreement with the date of the readvance maximum in the Western Forth Valley which may be inferred from the shoreline sequence in that area (Sissons, 1983a; Sutherland, 1984a). Dates on ice-transported molluscs from South Shian and Balure of Shian (Peacock *et al.,* 1989) have also indicated that the Creran glacier reached its maximum late in the stadial, possibly as late as 10,000 BP.

The timing of the retreat and final disappearance of the glaciers has been studied using both pollen analysis and radiocarbon dating of the basal sediments in enclosed basins in the deglaciated areas (Walker and Lowe, 1981, 1982; Lowe and Walker, 1981, 1991). Radiocarbon dating of such deposits has produced inconsistent results (Sutherland, 1980; Walker and Lowe, 1980) but may suggest that deglaciation of the west Highland icefield occurred between 10,600 BP and 10,200 BP (see Mollands and Kingshouse). A complementary approach to the establishment of the chronology of ice decay has been to examine the pollen assemblages from basal sediments in enclosed basins within the readvance limits. Such assemblages, when compared with those of regional pollen zonations for the end of the stadial and the early Holocene imply progressive ice retreat (Lowe and Walker, 1981, 1991; Walker and Lowe, 1985). These studies indicate that final deglaciation in the south-west Grampians, on Mull and on Skye was somewhat earlier than the maximum of *Juniperus* (Juniper) pollen values in the early Holocene (see below). However, this widely recognized feature of pollen diagrams is not well dated itself (Tipping, 1987) and could be as late as 9800 BP, which may imply final deglaciation of Scotland in the early Holocene.

The greatest extent of Loch Lomond Readvance glaciers was in the west Highland icefield (Figure 2.8) (Sissons *et al.,* 1973; Sissons, 1980b; Thorp, 1986, 1991a; Gray and Coxon, 1991) and significant subsidiary icecaps and icefields existed on Mull (Gray and Brooks, 1972), Skye (Walker *et al.,* 1988; Ballantyne, 1989a; Ballantyne and Benn, 1991) and in the south-east Grampians (Sissons and Grant, 1972; Sissons, 1974b). In addition, numerous smaller glaciers developed in climatically favourable localities in the northern Highlands (Sissons 1977a), the eastern Grampians (Sissons, 1972a, 1979b and the Southern Uplands (Sissons, 1967a; Cornish, 1981). Over 200 individual ice bodies have been identified (Sutherland, 1984a). Such a pattern probably reflects a combination of climatic factors and the form of the topography that controlled glacier initiation and subsequent development (Sissons and Sutherland, 1976; Payne and Sugden, 1990a, 1990b).

The limit of the readvance is marked in many areas by prominent end moraines. These may occur as massive accumulations of ice-pushed drift as at Menteith (see Western Forth Valley) and South Shian and Balure of Shian, as partly stratified deposits laid down in a glaciolacustrine environment (see Gartness) or as cross-valley ridges formed in a similar environment (see Glen Roy and Coire Dho), as bouldery ridges fronting corrie glaciers (see Lochnagar and the Cairngorms or as massive ridges of ice-transported debris (see Cnoc a'Mhoraire, Tauchers and An Teallach).

The Loch Lomond Readvance has also left a significant imprint on the landscape of many Highland valleys in the form of hummocky moraines; possibly the most outstanding example occurs at the Coire a'Cheud-chnoic in Glen Torridon. Another notable feature of the readvance was the creation of ice-dammed lakes as at Coire Dho (Sissons, 1977b), possibly (Ballantyne *et al.,* 1987) Achnaseen (Sissons, 1982a), Loch Tulla (Ballantyne, 1979) and, most famous of all, at Glen Roy (Jamieson, 1863, 1892; Sissons, 1978, 1979a, 1979c, 1981d; Peacock and Cornish, 1989). At all these localities former lake shorelines occur, but they are most clear and extensive at Glen Roy, where it has been demonstrated that they were cut mainly in rock and were faulted or warped, possibly due to rapid isostatic adjustments near the readvance limits (Sissons and Cornish, 1982a, 1982b). Sissons (1979a, 1979c) has also reconstructed the complex changes in lake drainage and river terrace development that occurred as the ice retreated from its maximum in Glen Roy and Glen Spean and hence has argued that there was one major and several more minor *jökulhlaups* (floods) during this deglaciation sequence. He suggested that a large fan of sand and gravel at Fort Augustus was formed by the major *jökulhlaup* and that the volume of water released was sufficient to raise the level of Loch Ness by several metres (see Dores) and also to build a large alluvial fan at Inverness blocking much of the Beauly Firth (Sissons, 1979c, 1981c).

The variations in the size of the glaciers and the altitude of their equilibrium lines allow inferences to be made about the climate during the period of glacier expansion. Mean summer temperatures of 7–9°C below those of the present have been deduced, and the pattern of precipitation was one of heavy snowfall in the western Highlands and along the Highland boundary, with markedly lower snowfall in the central and eastern Grampians and in the north of the country (Sissons, 1974b, 1980b; Sissons and Sutherland, 1976).

Terrestrial non-glacial environments

The severe climatic conditions of the stadial were reflected in the plant communities, as revealed by pollen analysis. A tundra-type vegetation occurred at low altitudes, with widespread presence of plants typical of unstable ground, although many of the plant communities that existed during this period appear to have no direct modern analogues (Walker, 1984b). Very high values of *Artemisia* pollen have been particularly noted in the eastern central Grampians, with declining frequencies to the west and south (see Abernethy Forest, Loch Etteridge, Stormont Loch and Loch an t-Suidhe) (Birks and Mathewes, 1978; Macpherson, 1980; Tipping, 1985). As certain species of *Artemisia* are chionophobus, the pollen values of that taxon have been interpreted as an indicator of precipitation during the stadial (Walker, 1975b; Birks and Mathewes, 1978; MacPherson, 1980; Lowe and Walker, 1986a). As such, the reduced levels of precipitation implied for the eastern central Highlands in general and Strathspey in particular correspond to the pattern of precipitation inferred from the distribution of contemporaneous glaciers (Sissons, 1980b). Certain pollen profiles also show variations in *Artemisia* pollen values during the stadial, with an early phase of low frequencies followed by a later phase of significantly higher frequencies (see Stormont Loch). Following the above line of argument, this change in *Artemisia* values has been interpreted as reflecting an early wet phase followed by a drier period (Caseldine, 1980a; MacPherson, 1980; Tipping, 1985, 1991a).

In southern Scotland knowledge of the stadial environment has been augmented by study of fossil Coleoptera (see Bigholm Burn), from which summer temperatures about 7°C below those of the present have been inferred (Bishop and

Coope, 1977; Atkinson *et al.,* 1987).

Under the severe conditions, non-glacial geomorphological processes were particularly active. The most dramatic landforms produced at this time were in the mountains, where a number of protalus ramparts formed (Ballantyne and Kirkbride, 1986), including the remarkable feature on Baosbheinn in Wester Ross (Sissons, 1976c) which may have been partly formed by a landslip (Ballantyne, 1986a). The distribution and altitude of protalus ramparts in the Highlands shows a pattern remarkably similar to that of the contemporaneous glacier equilibrium line altitudes; they are lowest in the west of the country and increase in altitude towards the east (see the Cairngorms) (Ballantyne, 1984; Ballantyne and Kirkbride, 1986). It may be inferred that the precipitation variations that were a major control on the glacier distribution also influenced the formation of protalus features. The protalus ramparts also give information about the rate of rockfall activity during the stadial (Ballantyne and Kirkbride, 1987), from which it is apparent that this process operated at a rate that is approximately an order of magnitude greater than at the present time and equivalent to some of the highest rates observed anywhere in the world today. Talus slopes formed during the Lateglacial are also of a size and maturity quite distinct from those that have accumulated during the Holocene (see An Teallach) (Ballantyne and Eckford, 1984).

The intense rockfall activity also contributed to the development of a number of rock glaciers (see Beinn Shiantaidh and the Cairngorms), although the most spectacular of these, at Beinn Alligin in Wester Ross, is most probably the product of a landslip on to the surface of a decaying glacier (Sissons, 1975a; Ballantyne, 1987c). Landslip activity in general may have been much more intense during the stadial than during the late Holocene and Holmes (1984) has highlighted a relationship between a large proportion of such features and the limits of Loch Lomond Readvance glaciers. He suggested that raised porewater pressures in slopes adjacent to the glaciers played a crucial role in the weakening of the rock, with subsequent landslipping at some stage after ice retreat.

Mountain-top periglacial processes can be inferred to have been particularly active during the stadial from the mutually exclusive relationship in many mountain areas between Loch Lomond Readvance limits and large-scale periglacial features (see Lochnagar, Western Hills of Rum, An Teallach and the Cairngorms). Although the mountain-top detritus comprising the periglacial features may have been derived from periods of intense frost weathering during Late Devensian ice-sheet deglaciation or even, on certain summits, during the glaciation itself, the relict periglacial landforms that can be observed today were most probably fashioned during the Loch Lomond Stadial. The lowest altitude at which these features occur on mountains declines westwards and northwards, implying some form of regional climatic control on their formation (Ballantyne, 1984). The specific types of landform developed on any particular summit, however, are a product of the available topography and the type of regolith derived from frost weathering of the underlying bedrock, in addition to climatic variables (Ballantyne, 1984). Thus the summits underlain by granites (see the Cairngorms, Lochnagar, Western Hills of Rum and Ronas Hill) display different periglacial landforms from those underlain by schistose rocks (see Ben Wyvis and Sgùrr Mòr), with further differences on summits underlain by sandstones (see An Teallach and Ward Hill).

At low altitudes, slope processes resulted in the destruction of the soil profiles that had developed during the preceding interstadial. In a number of areas interstadial pears or organic remains have been found underlying slumped or soliflucted material at the base of drift-covered slopes (see Bigholm Burn) (Donner, 1957; Gray, 1975b; Dickson *et al.,* 1976; Clapperton and Sugden, 1977). In enclosed basins the stadial is marked by a distinct layer of mineral sediment washed in from the surrounding slopes and frequently containing low concentrations of organic detritus derived from the interstadial soils (Lowe and Walker, 1977, 1986a; Pennington, 1977a; Walker and Lowe, 1990). River activity at low altitudes also appears to have been enhanced during the stadial, with large alluvial fans being built across interstadial sediments in the Lochwinnoch Gap (Ward, 1977), at the foot of the Ochil Hills (Kemp, 1971) and near Corstorphine in Edinburgh (Newey, 1970; Sissons, 1976b). A number of fossil frost wedges at low altitudes have been assigned to the stadial, suggesting permafrost conditions and mean annual temperatures of below –5°C (Rose, 1975; Sissons, 1977a, 1979e), but as none of these has been unequivocally dated to the stadial, climatic inferences must remain tentative.

Marine environments and sea-level change

The severe climate of the stadial was also registered in the contemporaneous marine fauna. Boreal species present during the interstadial disappeared and were replaced by high-arctic species such as *Portlandia arctica* (Gray) (Peacock *et al.,* 1978; Peacock, 1981b, 1983a, 1987, 1989b; Graham *et al.,* 1990). Nearshore marine summer temperatures of more than 10°C below present levels are implied. In offshore sediments, recognition and correlation of deposits from the stadial have been aided by the occurrence of shards of volcanic ash (the Vedde Ash) which has been dated to approximately 10,600 BP (Long and Morton, 1987; Graham *et al.,* 1990). This ash layer is considered to be derived from a volcano in Iceland and has been found widely along the west coast of Norway (Mangerud *et al.,* 1984) but not on-land in Scotland. Its occurrence in marine sediments off both the east and west coasts of Scotland may, however, have been due to redistribution by sea ice, depending on the time of the year when the volcano was in eruption.

In the Forth valley and along the Firth of Forth, a distinct erosional surface has been identified cutting across Lateglacial marine sediments, till and bedrock. It is immediately overlain by early Holocene sediments (Sissons, 1969, 1976a) and has been accordingly inferred to have been formed towards the end of the Lateglacial. The inner margin of the surface marks a clear, isostatically tilted shoreline termed the Main Lateglacial Shoreline. Its distinctive erosional nature in a low-energy environment which was normally characterized by fine-grained estuarine deposition, together with its stratigraphic position, led Sissons (1976b) to conclude that the feature was formed mainly during the severe climatic conditions of the stadial. A similar surface has been found in the inner Beauly Firth (see Barnyards) (Sissons, 1981c; Firth, 1984).

Around the coasts of the south-west Highlands and the neighbouring islands there is also a distinctive marine erosional feature (see Isle of Lismore, Glenacardoch Point and West Coast of Jura), the Main Rock Platform (Gray, 1974a; 1978a; Dawson, 1980b, 1988a; Gray and Ivanovich, 1988), which is tilted to the south and west in conformity with the local pattern of isostatic deformation. It also has a gradient similar to that of the Main Lateglacial Shoreline of the east coast. These similarities led Sissons (1974d) to conclude that the Main Rock Platform had also been produced by particularly active frost erosional processes in the intertidal zone. Such a hypothesis received support from those who subsequently worked on the platform (Gray, 1978a; Dawson, 1980b; Sutherland, 1981b) and the concept of rapid erosion of bedrock in a littoral environment during the stadial appeared to be confirmed by evidence from the Parallel Roads of Glen Roy. More recently, however, certain evidence seems to suggest that the platform may have been in part inherited from a pre-existing feature or features (Browne and McMillan, 1984; Gray and Ivanovich, 1988), thus diminishing (though not excluding) the need to postulate such active erosional processes during the stadial.

It is possible that the ice masses that developed during the readvance were of sufficient magnitude to re-depress the crust isostatically (Sutherland, 1981b; Firth, 1986, 1989c; Boulton *et al.,* 1991), and in the Western Forth Valley the maximum of the readvance coincides with a marine transgression which followed the formation of the Main Lateglacial Shoreline. At the peak of this transgression a distinct shoreline (the High Buried Shoreline) was formed seawards of the Menteith Moraine, but it is not found within the limits of the readvance (Sissons, 1966, 1983a). This rise of sea level, which was at least 8 m, coincided with the readvance maximum and is most easily explicable as the result of renewed isostatic depression, as there does not appear to be a significant coincidental eustatic rise in sea level that would otherwise explain it. Using a calculation of shoreline gradient against age, it is possible to infer that the High Buried Shoreline was formed between 10,000 BP and 10,500 BP (Sutherland, 1984a), thus placing a limit on the time of the maximum of the readvance in the Forth valley.

In summary, the Loch Lomond Stadial was a period of particularly marked environmental change and it was also the last period when glaciers existed in Scotland. In contrast to the preceding interstadial and subsequent early Holocene, the stadial demonstrates the impact of a severe, albeit short-lived (in geological terms), climatic deterioration and shows the sensitive and marginal nature of the Scottish environment to such a change. An interesting footnote is the possibility that Man was present in Scotland at this time, as has been inferred from a collection of reindeer antlers in the Creag nan Uamh caves in Sutherland (Lawson and Bonsall, 1986a, 1986b).

The Holocene

The boundary between the Late Devensian and the Holocene is conventionally placed at 10,000 BP, although the actual climatic amelioration at the end of the Loch Lomond Stadial may have occurred somewhat prior to this. The change in climate coincided with the return once more of North Atlantic Drift waters to the Scottish coasts. Peacock and Harkness

(1990) have indicated that the warming occurred in two phases, with a period between approximately 10,100 BP and 9600 BP when marine temperatures were similar to those during the Lateglacial Interstadial, that is 2–3°C below the present level. The greater part of the warming that occurred slightly prior to 10,100 BP may have taken only 40 years (Peacock and Harkness, 1990). Similar extremely rapid temperature changes have been inferred from fossil coleopteran assemblages, and atmospheric temperature changes could have been of the order of 1°C per decade (Coope, 1977; Atkinson *et* a/., 1987).

The brief period of 'interstadial' conditions at the beginning of the Holocene that has been deduced from the nearshore marine faunas has not been noted in the vegetational record as inferred from pollen analytical studies, possibly because of the slow response time of plants to the major climatic amelioration. However, at two sites on Mull (see Gribun) and on Skye, which have long early Holocene sequences and hence permit very detailed analysis of vegetation changes, brief vegetation 'revertance' phases have been noted in the early Holocene (Dawson *et al.,* 1987a; Lowe and Walker, 1991).

Vegetational development

Immediately after the amelioration of climate, the vegetation throughout most of Scotland was dominated by open-habitat taxa, these being succeeded, as during the early part of the Lateglacial Interstadial, by shrub and scrub vegetation. Unlike the opening of the earlier interstadial, however, when an initial mild phase was soon followed by deterioration in climate with temperatures declining to somewhat lower than at present, in the Holocene the mild climate was maintained and hence the pattern of vegetational development was quite distinct from that of the interstadial. The dwarf shrub and scrub phase saw successive dominance in much of the country of *Empetrum* and then *Juniperus* (this phase being particularly marked). Subsequently there followed expansion of mixed deciduous woodland.

The pattern and timing of the spread of tree species into Scotland during the early to middle Holocene may broadly be conceived in terms of expansion from the south, but when viewed in detail this is clearly an oversimplification (Huntley and Birks, 1983; Birks, 1989). *Betula* was the first tree to spread across the country, which it did between 10,000 BP and 9000 BP. Earliest expansion was in the south-east and east, which may partly be explained by spreading from areas of dry land in the south and central North Sea Basin. *Corylus* (hazel) shows a completely different spreading pattern, with very early, rapid migration along the western coastal fringe and the Inner Hebrides, the northern coast being reached by 9500 BP. Thereafter, by 9000 BP, it had expanded across southern and central Scotland and, by 8500 BP, into the central Highlands and the north-east. The initial spread along the west coast may have been due to water transport of hazel nuts combined with the occurrence (by analogy with today) of infrequent frosts along the western coastal fringe.

Ulmus (elm) and *Quercus* (oak) both showed a broadly similar south-to-north spreading pattern. *Ulmus* arrived first, expanding across southern and central Scotland and into the Highlands between 9000 BP and 8500 BP. This rapid expansion, at around 500 m a^{-1} thereafter slowed and the tree did not reach the north-east until 8000 BP and Caithness until 6500 BP. *Quercus* spread through southern and central Scotland and into the south-west Highland coastal fringe between 8500 BP and 8000 BP and had reached its northern limits in southern Skye and on the southern shore of the Moray Firth by 6000 BP.

The spread of *Pinus* (pine) had perhaps the most curious pattern (Bennett, 1984; Birks, 1989). Its earliest occurrence was in the northern Highlands at around 8500–8000 BP (Birks, 1972b) (see Loch Maree and Loch Sionascaig) and there was subsequent spread into the eastern Highlands at around 7500 BP (see Abernethy Forest) and throughout much of the central Grampians by 5000 BP. Pine was not a significant element in the woodlands at any time through almost the whole of south-central Scotland and the south-west Highland coastal fringe (Bennett, 1984), an exception occurring in the western Southern Uplands (Birks, 1972a, 1975) (see Loch Dungeon), where expansion from Ireland at around 7500–7000 BP occurred (Birks, 1989). This unusual pattern is as yet unexplained, but the northern Scots Pine may be genetically distinct from that which occurred in southern Britain (Kinloch *et al.,* 1986), suggesting expansion from a different source area.

Alnus (alder) also exhibits an unusual pattern of spread (Birks, 1989; Bennett and Birks, 1990). It expanded through Scotland between 9000 BP and 5000 BP, but its occurrence was sporadic in both space and time, apparently due to its specific wet, mildly basic habitat requirements.

By the later part of the middle Holocene, the Scottish forests had reached their greatest extent and degree of diversity (Figure 2.12) (Bennett, 1989). In the very north-west of Sutherland and the northern and western island fringes there was little forest development, but birch and pine stumps have been recovered from peat profiles (see Garths Voe), suggesting scattered occurrences of woodland in favoured locations during the middle to early part of the late Holocene. After 5000 BP, a change to a cooler, moister climate together with human impact led to a reduction in tree cover and a corresponding expansion of blanket peat, heaths and grasslands.

The impact of Man was first noted slightly prior to 5000 BP, with a marked decline in elm pollen throughout the country (see Din Moss, Black Loch, Loch Cill an Aonghais, Loch Ashik and Loch Maree). This decline was accompanied or followed by increases in grasses, heaths and species that are considered typical of cultivation. In more recent times, sites such as Black Loch (Whittington *et al.,* 1991a), which have rapid sedimentation rates, have allowed identification of the timing of the introduction of exotic species as well as changes in agricultural practice during historical times. Of particular interest are the detailed studies that have been carried out using pollen and diatom analyses on the acidity of waters in enclosed basins. The most detailed such study, at Round Loch of Glenhead in the western Southern Uplands (Jones *et al.,* 1989), has reconstructed the acidity of the basin waters since the Lateglacial and demonstrated clearly the impact of industrial pollution of the area in the last 200 years.

Sea-level change

Holocene changes of sea level were quite different along those parts of the coast strongly influenced by glacio-isostatic effects of the last ice-sheet (and probably also by the Loch Lomond Readvance ice-field) and the more peripheral regions where isostatic effects were minimal. In these latter regions, the world-wide glacioeustatic rise of sea level during the early to middle Holocene was the dominant influence.

The most detailed studies of early Holocene sea-level change have been made along the coasts and the estuaries of eastern Scotland. The isostatic effects in this region were greatest in the Western Forth Valley and near the head of the Beauly Firth (see Barnyards) and declined eastwards, being least in north-east Scotland (see Philorth Valley). In the areas most influenced by isostatic uplift, relative sea level in the early Holocene was initially high, with a prominent shoreline (Main Buried Beach) being formed in the Western Forth Valley (Sissons, 1966, 1972b) at around 9600 BP. This shoreline is probably also present in the Tay–Earn valley (see Carey) (Cullingford *et al.,* 1980) and in the Beauly Firth area (see Barnyards). Following the abandonment of this shoreline, sea level fell and a lower beach (Low Buried Beach) was formed at around 8600 BP. Then sea level again dropped, reaching its lowest level at between 8500 BP and 8300 BP in the Forth, Tay and Beauly Firth areas (Sissons and Brooks, 1971; Cullingford *et al.,* 1980; Haggart, 1989). Studies on the west coast of Scotland, along the Solway Firth and the Firth of Clyde, also suggest that this period was one of relatively low sea level, with peat formation at altitudes close to present sea level (see Newbie, Redkirk Point and Dundonald Burn) (Jardine, 1975, 1977, 1980b; Bishop and Coope, 1977; Sutherland, 1981b; Boyd, 1982).

The early Holocene fall of sea level in those areas near to the centre of isostatic uplift was a consequence of the rate of uplift exceeding the rate of world sea-level rise. Subsequently, however, this pattern was reversed and a transgression (the Main Postglacial Transgression) occurred everywhere around the Scottish coast. This relative sea-level rise resulted in the deposition of estuarine and marine sediments on top of early Holocene peats that had formed during the period of low sea level. Thus the progress of the transgression is marked by the time-transgressive nature of the overlap of the marine or estuarine sediments on to the terrestrial deposits. This process has been clearly registered at sites around the Solway Firth and the Firth of Clyde (see Redkirk Point, Newbie and Dundonald Burn), in the Western Forth Valley, along the Tay estuary and neighbouring parts of the east coast (see Carey, Silver Moss and Maryton) and at the head of the Beauly Firth (see Barnyards).

At a number of sites along the east coast the apparently simple stratigraphy described above is interrupted by a distinct thin bed of fine sand which cuts across the estuarine sediments, tapering out in the adjoining terrestrial peats. This bed contains marine diatoms and has been radiocarbon dated to around 7000 BP (see Barnyards, Maryton, Silver Moss and Western Forth Valley). It has been interpreted as either the product of a major storm surge in the North Sea Basin (Smith *et al.,* 1985a; Haggart, 1988b) or the result of a tsunami caused by a major submarine slide on the edge of the Norwegian continental shelf to the north of the North Sea (Dawson *et al.,* 1988; Long *et* al., 1989a). It is an important time-stratigraphic marker in the east coast marine and estuarine sediments, the variations in altitude of which, for example, provide precise information on the regional pattern of isostatic uplift in the last 7000 years (Long *et al.,* 1989a).

The maximum of the transgression appears to have been reached at different times in different parts of the country, being earliest in the Western Forth Valley (at around 6800 BP: Sissons, 1983a) and generally later in areas farther from the centre of isostatic uplift (see Silver Moss and Philorth Valley) (Smith *et al.,* 1983; Cullingford *et al.,* 1991). A particularly detailed study of the period close to the maximum of the transgression at Pitlowie in the Tay estuary (Smith *et ca.,* 1985b) has revealed that sea level, after rising rapidly to near its maximum altitude, stayed relatively stable for approximately 1000 years. This relatively stable period may be registered as a double peak to the transgression in areas of greater isostatic uplift such as Loch Lomond (see South Loch Lomond), which was an arm of the sea at this time (Dickson *et al.,* 1978; Stewart, 1987; Browne and McMillan, 1989). A par ticularly prominent depositional shoreline, the Main Postglacial Shoreline, was formed at this time and it has been shown to be isostatically tilted away from a centre of uplift in the southwest Highlands (Sissons, 1983a; Cullingford *et al.,* 1991) (see Western Forth Valley, West Coast of Jura, Munlochy Valley and Dryleys).

The phase of transgression was the result of a rapid rise in world sea level that terminated at around 6000 BP (by which time both the Scandinavian and the North American ice-sheets had melted) and, as isostatic uplift was still continuing around most of the Scottish coast, albeit at a slower rate than in the early Holocene, a fall in sea level ensued. This fall is not well dated, but a series of isostatically tilted shorelines were formed (see Munlochy Valley, Dryleys and West Coast of Jura), possibly during halts in the general regression.

In the more peripheral areas there is little direct information on sea levels during the early Holocene, which appear everywhere to have been much lower than at present. By the middle Holocene, relative sea level was rising to close to the present level in the Outer Hebrides (see Borve) (Ritchie, 1966, 1985) but was apparently significantly lower than this in Shetland (Hoppe, 1965). The middle to late Holocene in these areas appears to have been a time of relatively stable or slightly rising sea level and, in contrast to the glacio-isostatically affected areas, no Holocene raised shorelines have been found.

Geomorphological processes

The establishment of soil and vegetation covers during the early Holocene together with a milder climate resulted in a marked diminution in the intensity of geomorphological activity compared with that of the Loch Lomond Stadial. However, the Late Devensian left a legacy in the form of large volumes of unconsolidated debris and oversteepened slopes. Consequently, during the early Holocene there appears to have been a considerable number of slope failures (Watters, 1972; Holmes, 1984; Ballantyne, 1986d, 1991c) as well as frequent reworking of the available debris by streams and debris flows to produce, respectively, river terraces and debris cones in many Highland valleys (see Eas na Broige and Glen Feshie). This activity appears to have slowed during the middle to late Holocene, but there has been a resurgence in the last 200–300 years (cf. Ballantyne, 1991c), possibly due to the effect of the climatic deterioration termed the 'Little Ice Age' (Brazier and Ballantyne, 1989) and possibly also under the influence of increased grazing pressure in the upland areas (Innes, 1983b; Brazier, 1987; Brazier *et al.,* 1988). Apart from in a few cases, however, the precise causes remain uncertain (Innes, 1983a; Ballantyne, 1991a).

Periglacial activity has continued on hill tops and mountain summits throughout the Holocene (see Ward Hill, Western Hills of Rum, An Teallach, Sgiirr Meg·, Ben Wyvis, the Cairngorms and Tinto Hill), but, as at lower altitudes, the intensity of the processes has been markedly less than during the Loch Lomond Stadial (Ballantyne, 1987a, 1991c). Small-scale forms are typical of the Holocene and they show the same dependence of type on the underlying regolith as described above for the large-scale periglacial features formed during the colder periods. There also occurs a marked diminution northwards and westwards in the altitude at which given features form and, as noted above for the larger examples, this again implies an overall regional climatic control.

Where weathering of the underlying bedrock has released considerable quantities of sand, certain hills have areas covered in sand sheets (see Ronas Hill, Ward Hill and An Teallach) which contain organic layers. Dating of the latter suggests that during the early Holocene there was a period of sand accumulation followed, during much of the middle and late Holocene, by relative inactivity. However, in recent centuries there has been reactivation of sand movement, possibly due to grazing pressure (Ball and Goodier, 1974; Ballantyne and Whittington, 1987). As vegetation cover also has a major role in stabilization of the regolith, reduction in vegetation cover due to grazing pressure and/or climatic deterioration seems to have been responsible for a recent increase in the area of patterned ground on Tinto Hill (Miller *et al.,* 1954) and rapid solifluction on the Fannich mountains in the last few centuries (Ballantyne, 1986c).

Conclusion

There has been an upsurge in studies of the Quaternary during the last 20 years and, with the application of dating techniques as well as micropalaeontological analyses to both the marine and the terrestrial records, the foundations have been laid for a greatly improved understanding of the Earth's climatic system and the factors that govern climatic change. It has become apparent that relatively minor changes in the Earth's orbital parameters are amplified by 'feedback loops' in the atmosphere–ocean–cryosphere–biosphere system to produce the major climatic changes that have characterized the Quaternary. The North Atlantic oceanic circulation is a particularly important element in the Earth's climatic system, for not only is that ocean the source of the greater part of the precipitation that nurtured the principal northern hemisphere ice-sheets during ice ages but the North Atlantic is also, at present, the principal area of formation of oceanic deep water. Formation of deep water releases large amounts of heat to the North Atlantic surface waters and helps maintain the equable climate experienced in north-west Europe. During glaciations, however, deep water formation is retarded or halted, with consequent impact on the climate of the surrounding land masses.

The position of Scotland, jutting into the North Atlantic and with the west coast washed by the North Atlantic Drift, has therefore produced a climate and natural environment that are particularly dependent on and sensitive to changes in the circulation of the North Atlantic. Consequently, during the Quaternary Scotland has at different times been completely covered by ice, partially glaciated with accompanying intense non-glacial geomorphological activity beyond the ice margins, or, as at present, has experienced a mild, temperate climate. Coincident with the changes between the different environments there have been great fluctuations in sea level around the Scottish coasts and on the neighbouring continental shelves from more than 40 m above to less than 120 m below present level.

The challenge presented by these environmental changes is to establish their pattern and chronology from the evidence around Scotland and to understand how the different parts of the environmental system respond. In a global context, Scotland is significant in certain respects. First, the climatic record of its onshore and offshore deposits should mirror the changes in the circulation of the North Atlantic. Second, the relatively small size of the ice masses that developed in Scotland implies that they would have had a faster response time to climatic change than the Scandinavian and North American ice-sheets. Consequently, it may be anticipated that the glaciers and accompanying environmental fluctuations in Scotland were a sensitive monitor of global climatic change, particularly those changes of higher frequency and lower magnitude. At a national scale, knowledge of the mechanisms controlling change in the natural environment provides the necessary baselines against which human impacts on the environment can be assessed and differentiated from 'natural' environmental change. Also, by providing a temporal perspective, it allows a better understanding of the diversity of the natural environment and the potential rates of change arising from natural perturbations or human interference.

As is apparent from this review of current understanding of the Scottish Quaternary, much work remains to be done before even a reasonably complete record of Quaternary events is established. However, the existing information has offered important insights, at a variety of scales, into environmental stability and response times as well as the rates of change possible in different parts of the environment once thresholds are crossed and change initiated. The present review therefore concludes with three examples of response to environmental change which highlight the importance of thresholds in the natural environment and the small changes that may be sufficient to switch natural systems rapidly from stable to unstable states or from one level of dynamic equilibrium to another.

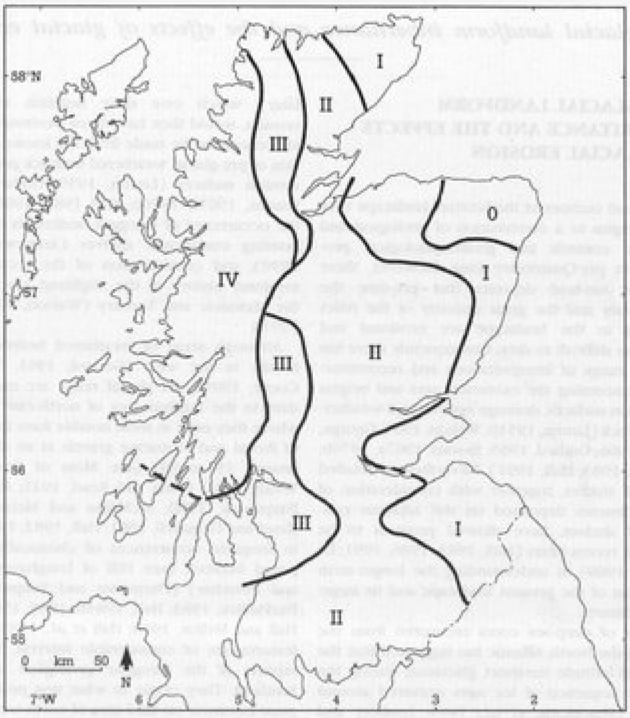
During the Quaternary, the evidence of both temperature and ice-sheet volume changes derived from deep-sea sediments indicates that Scotland is likely to have been glaciated on numerous occasions. However, in this review evidence has been found for only four (less certainly, six) major expansions of ice-sheets across the lowlands and on to the adjacent continental shelves. Even accepting that the record is incomplete, the implication is that major ice-sheet expansion is relatively uncommon. In addition, it has been suggested above that the expansion of the Late Devensian ice-sheet from the Highlands to its maximum and subsequent retreat to within the Highlands once more may only have taken approximately 10,000–12,000 years, with average rates of advance and retreat of the ice margin of 30 m a^{-1}. Such rates are of a similar order of magnitude to those observed at the North American and Scandinavian ice-sheet margins and, as noted by Ehlers *et al.* (1991), imply ice velocities comparable to the fastest Antarctic and Greenland ice streams, which are difficult to comprehend for an ice-sheet as a whole. In addition, almost the whole of the retreat of the last Scottish ice-sheet from its maximum position to within the Highlands occurred during a time of cold climate. The behaviour of the Scottish ice-sheet therefore appears to be one in which it is confined within the mountain zone until a certain threshold is crossed, whereupon it expands and then retreats rapidly. The ice-sheet modelling experiments of Payne and Sugden (1990b) may provide insight into one of the mechanisms responsible for this behaviour, for they have observed that the intra-mountain basins (such as Rannoch Moor) provided a threshold for ice-sheet expansion. They noted that only a very slight change was necessary in the climatic parameters controlling glacier development for a stable intra-mountain ice-field to change to a rapidly growing ice-sheet and that this occurred when the infra-mountain basins changed from being areas of ablation to areas of accumulation.

A quite different example is the rates of change in the biosphere that resulted from the major climatic ameliorations during the Lateglacial and at the opening of the Holocene, which have been found to be quite distinct between different species. The climatic changes, of the order of 5–10°C or more increase in mean annual temperatures, have been estimated to have taken place in only several decades and certain species of Coleoptera as well as of marine micro- and macro-faunas have been able to respond within these time periods. However, there was a delayed response by most plant species, not least because many have complex habitat requirements. At the opening of the Lateglacial Interstadial, however, climatic conditions similar to those of today were only maintained for 100–300 years before temperatures fell a few degrees. This decline was sufficient to preclude the spread of trees into Scotland during the interstadial, with the exception of tree birch in the south and east and possibly pine in favoured areas. In contrast, at the beginning of the Holocene the mild climate was maintained and created conditions that were suitable for trees to spread and through the southern and central parts of the country rates of expansion were very high, between 150 and 500 m a^{-1} on average. Towards the limits of their ranges in the north of the country expansion was much slower, typically less than 100 m a^{-1}. These rates of expansion are far higher than can be understood in terms of propagation directly from trees and additional mechanisms such as water or bird transport need to be invoked.

Finally, it is apparent from the studies of Holocene geomorphological processes that there is a very fine balance between stability and erosion in Scottish upland areas and that during the last 200–300 years this balance has been tilted in favour of erosion. This has been noted in increased debris-flow and alluvial fan activity, erosion of blanket peat, renewed activity of mountain-top aeolian sands and, at least locally, rapid mass movement of frost-derived debris on mountain summits. Upland blanket pears have accumulated during the Holocene at average rates of between 1.5 and 5 cm 100 a^{-1}. Today, almost throughout the upland areas, peat erosion, in places to depths of over 1 m, is occurring. This phase of erosion commenced 200–300 years ago and is apparently far more intense than occasional brief periods of erosion during the late Holocene (Stevenson *et al.,* 1990). The precise mechanism that has induced this erosion is not yet understood but may be the result of increased burning or grazing pressure or it may be a response to the climatic downturn known as the 'Little Ice Age'.

These three examples illustrate at both regional and local scales the rapidity with which the natural environment can change. All the changes are, or would have been, appreciable on a 'human' time-scale measured in decades. The causes of the changes were principally natural, although the role of human impact cannot yet be fully separated from natural changes in the third example. In all the examples there appears to be a rather narrow dividing line between an essentially 'stable' condition and rapid change. Future Quaternary studies will lead to elucidation of the factors involved and also a better understanding of landscape sensitivity, which is essential in developing sustainable environmental management and use of natural resources.

References

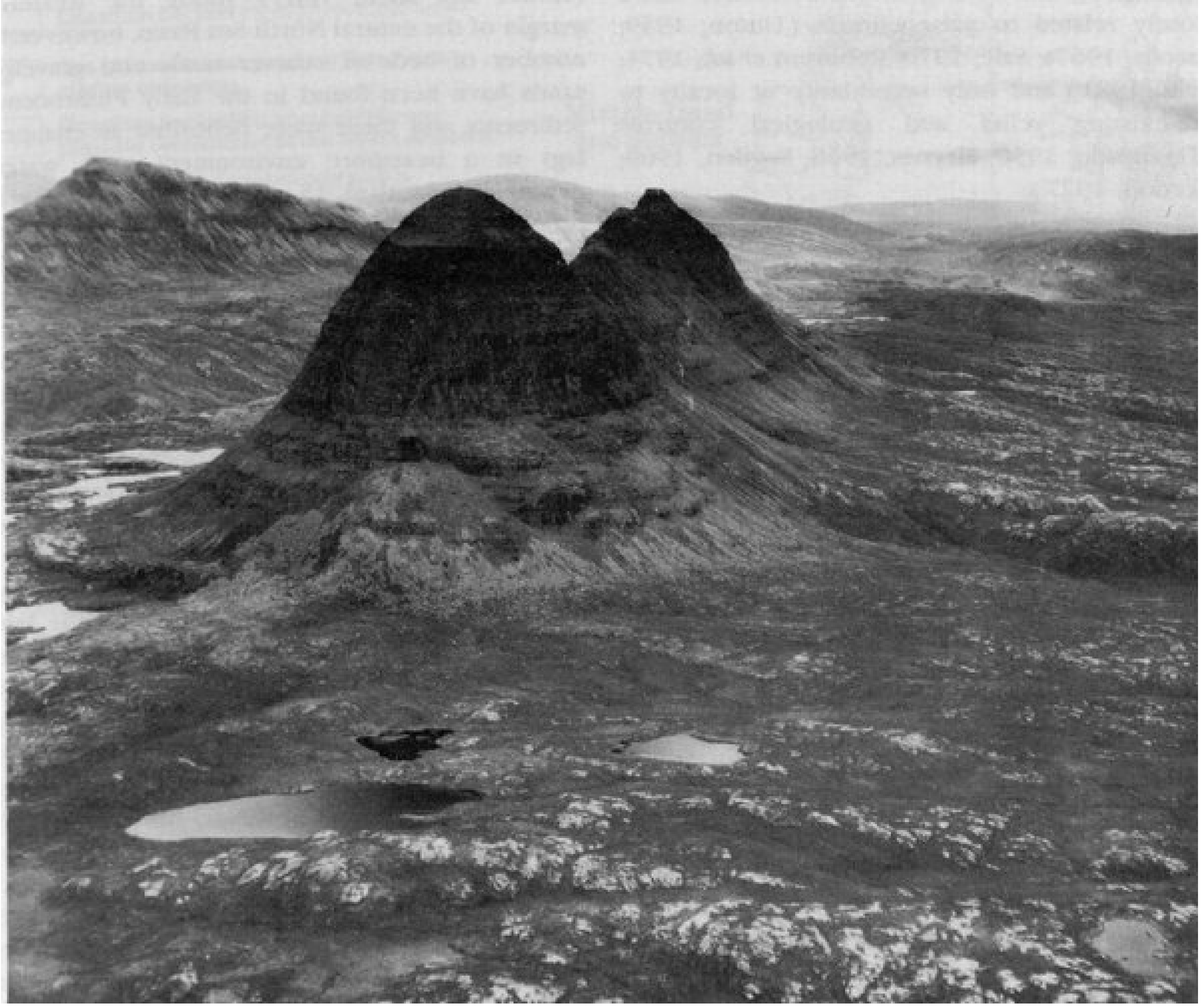


Zone	Lowlands	Uplands
O	No erosion Head on weathered rocks and slopes Outwash in concavities Rare occurrences of till on weathered rock	No erosion Outwash on valley floors Solifluction deposits on slopes Boulder fields and tors on divides
I	Ice erosion confined to detailed or subordinate modifications Concavities drift mantled but convexities may show some ice moulding Occasional roches moutonnées Ice-scoured bluffs in favourable locations	Ice erosion confined to detailed or subordinate modifications Suitable valley slopes ice steepened Entrenched meanders and spurs converted to rock knobs Interfluves still commonly Zone O
II	Extensive excavation along main flow-lines so that concavities may be drift free or floored by outwash or post-glacial deposits. Isolated obstacles may be given ovoid or cutwater forms if of soft rock, or crag-and-tail with associated scour troughs if hard Margins of larger masses converted to ice-scoured bluffs or planar slopes	Conversion of pre-glacial valleys to troughs common, but usually confined to those of direction concordant with ice flow Some diffluence; transfluence rare Interfluves may be Zone I or even Zone O, and separated from troughs by well-marked shoulders
III	Pre-glacial forms no longer recognizable but replaced by tapered or bridge interfluves with planar slopes on soft rocks, and by rock drumlins and knock-and-lochan topography on hard rocks	Transformation of valleys to troughs comprehensive giving compartmented relief with isolated plateaus or mountain blocks. Zone O may still persist on interfluves at sufficiently great heights
IV	Complete domination of streamlined flow forms even over structural influences	Ice moulding extends to high summits Upland surfaces given knock-and-lochan topography (sometimes of great amplitude at lower levels) Lower divides extensively pared or streamlined

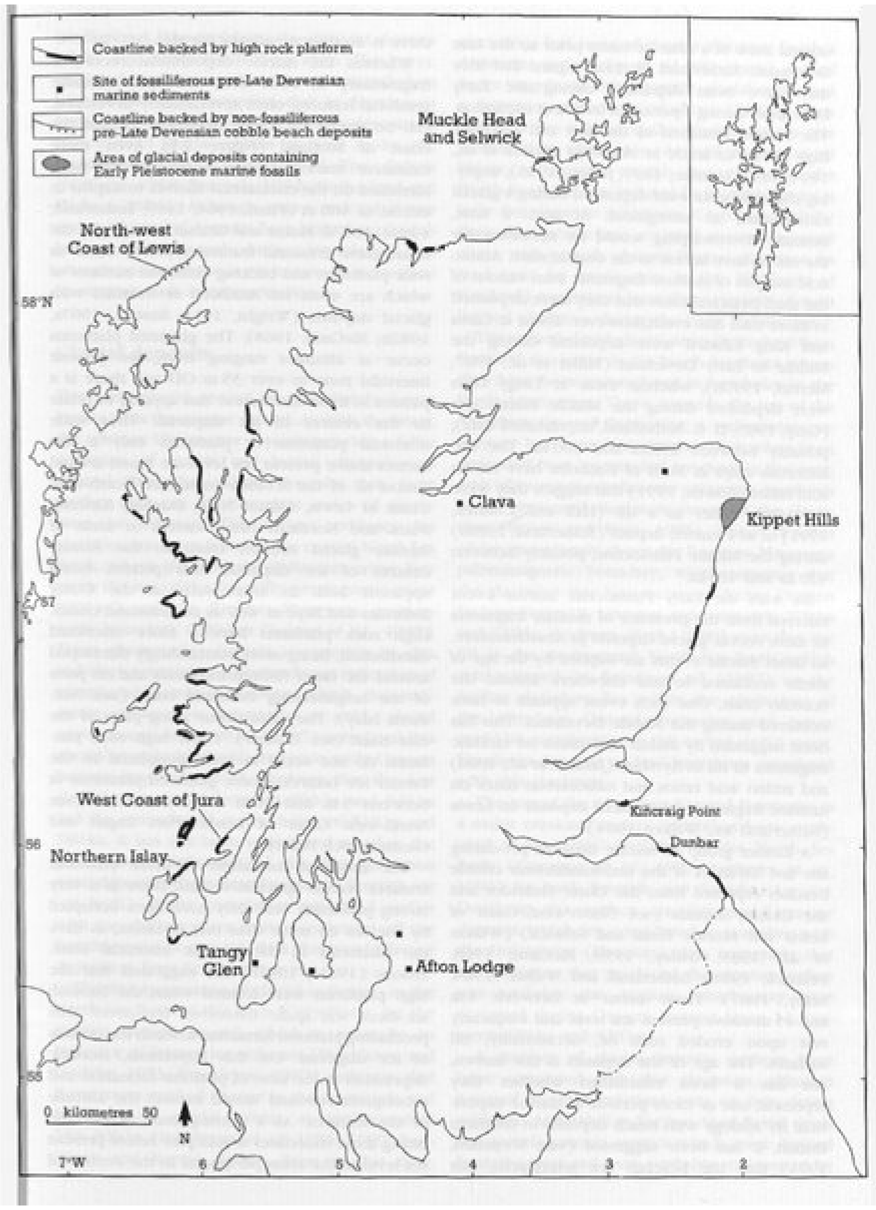
(Figure 2.1) Zones of glacial erosion (from Clayton, 1974).

(Figure 2.2) Distribution of rock basins On the Scottish mainland and neighbouring shelves.

(Figure 2.3) The western Cairngorms, with Ben Macdui (centre) viewed from the east, showing the diversity of the geomorphology of the eastern Highlands. Rounded plateau surfaces with tors (on Beinn Mheadhoin, lower left), solifluction lobes and weathered bedrock contrast strikingly with landforms of glacial erosion, including corries, the glacial troughs of Loch Avon (foreground) and the Lairig Ghru (middle distance), and the truncated spur of the Devil's Point (top, left of centre). Locally, parts of the plateau are ice scoured, as between Ben Macdui and Loch Avon. 'Hummocky moraine' can also be seen at the head of Loch Avon. (Cambridge University Collection: copyright reserved.)

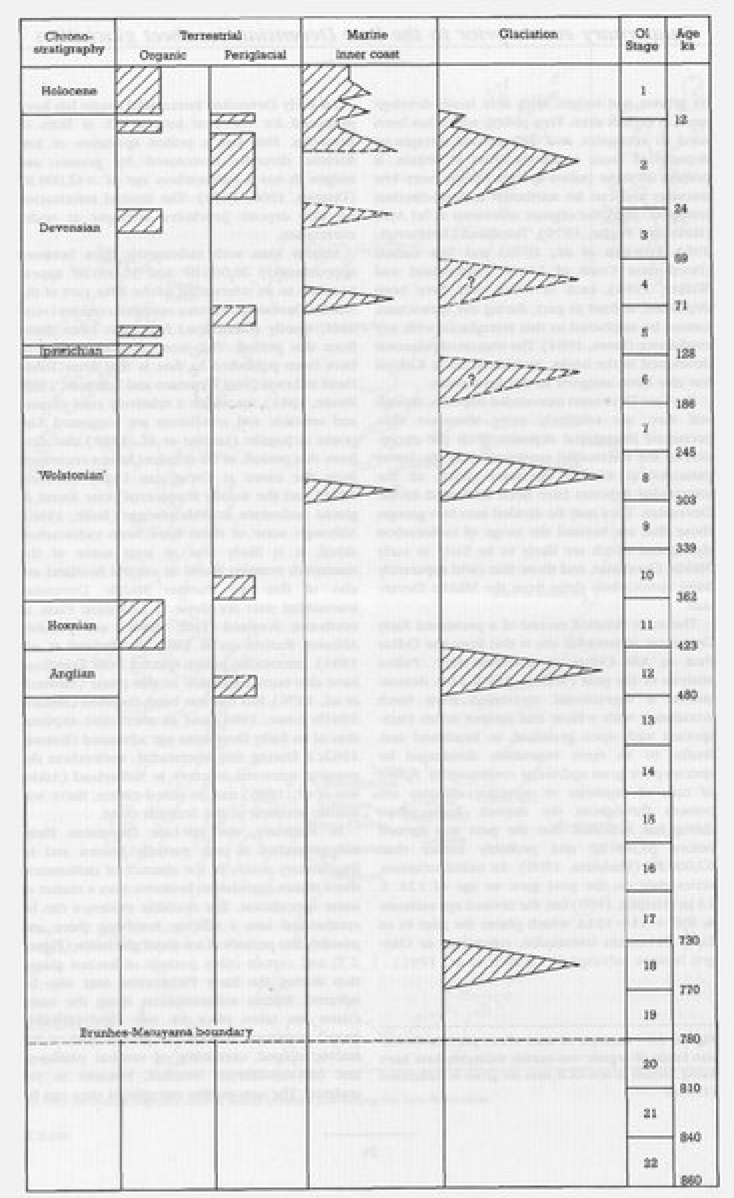


(Figure 2.4) Suilven, with Canisp in the background (left), shows the striking relationships between geology and relief in north-west Scotland. These isolated mountains of Torridonian sandstone rise above the intensively ice-scoured surface of Lewisian gneiss. (© British Crown copyright 1992/MOD reproduced with the permission of the Controller of Her Britannic Majesty's Stationery Office.)

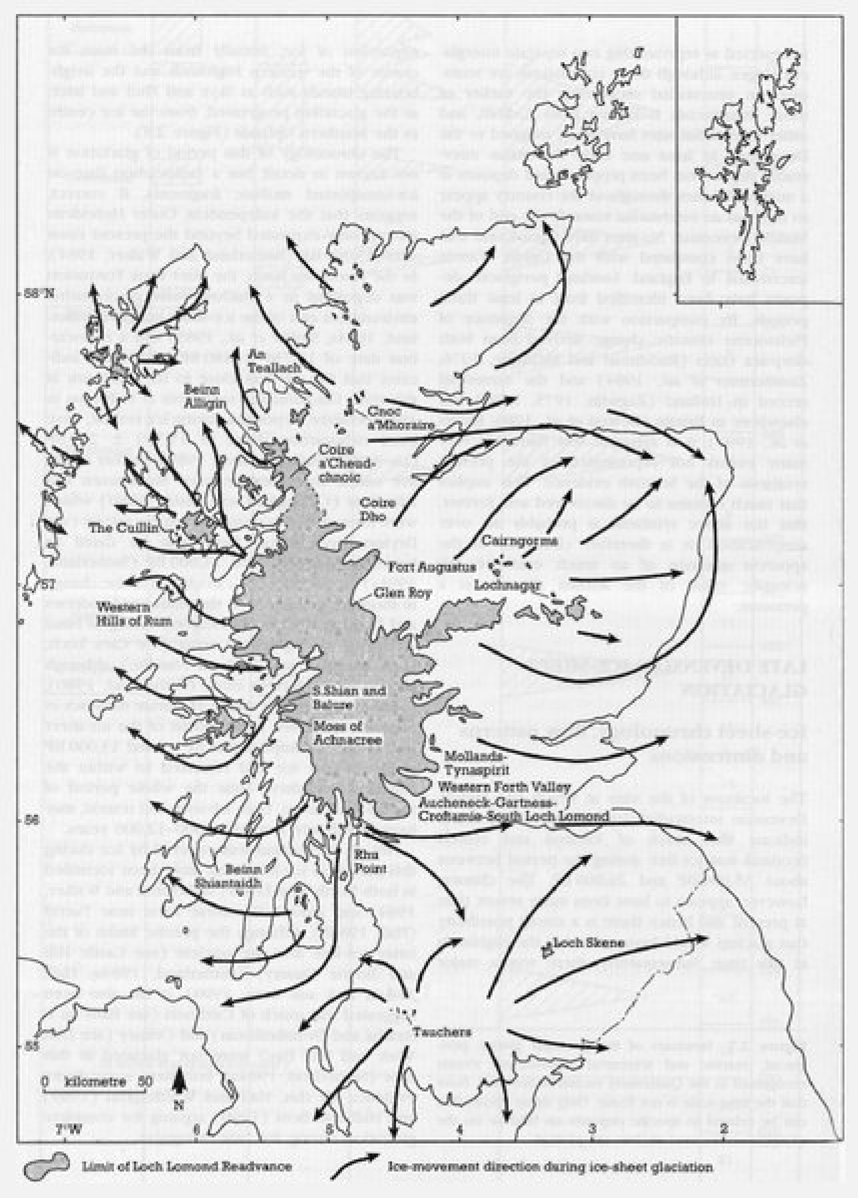


(Figure 2.5) Distribution of pre-Late Devensian marine deposits and high rock platforms.

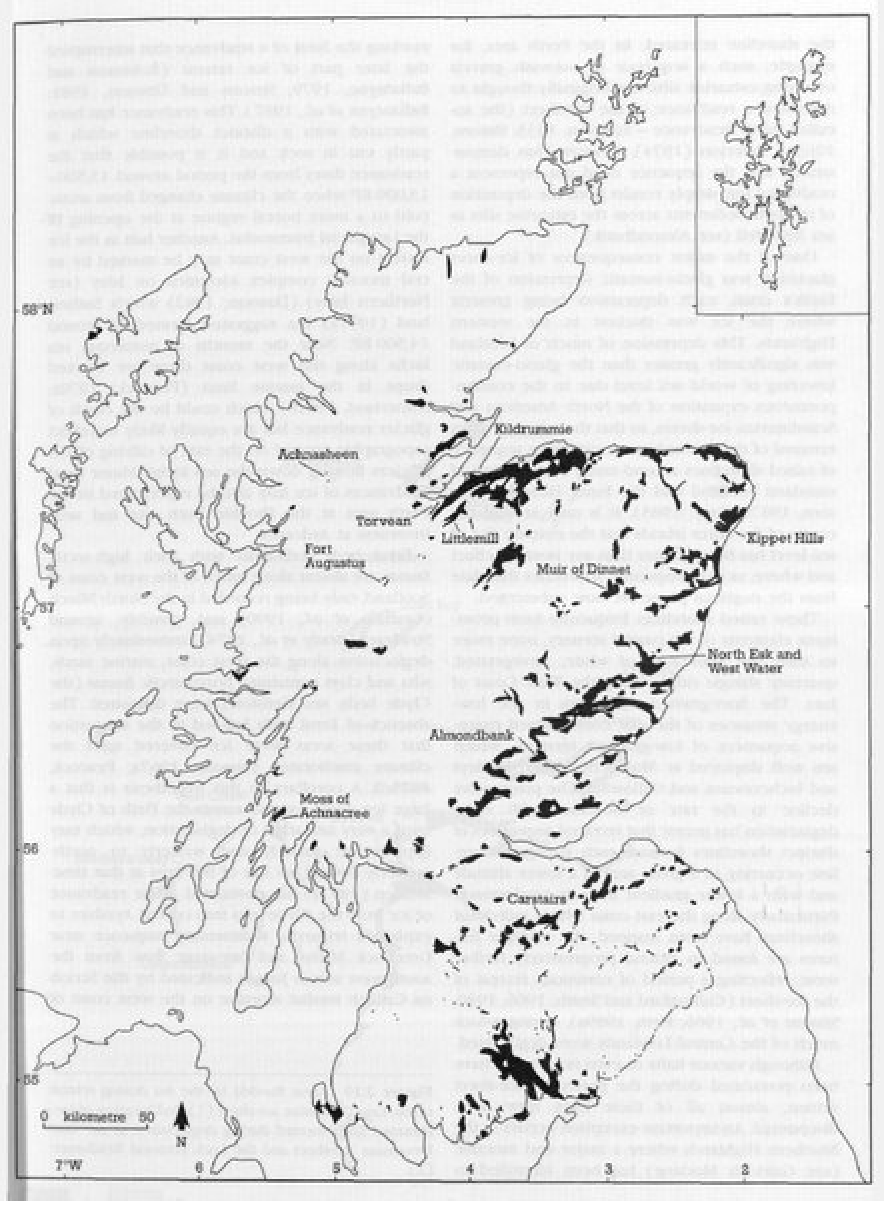
(Figure 2.6) Location of sites where pre-Late Devensian fossils or organic non-marine sediments have been found. Details of non-GCR sites are given in Sutherland (1984a).



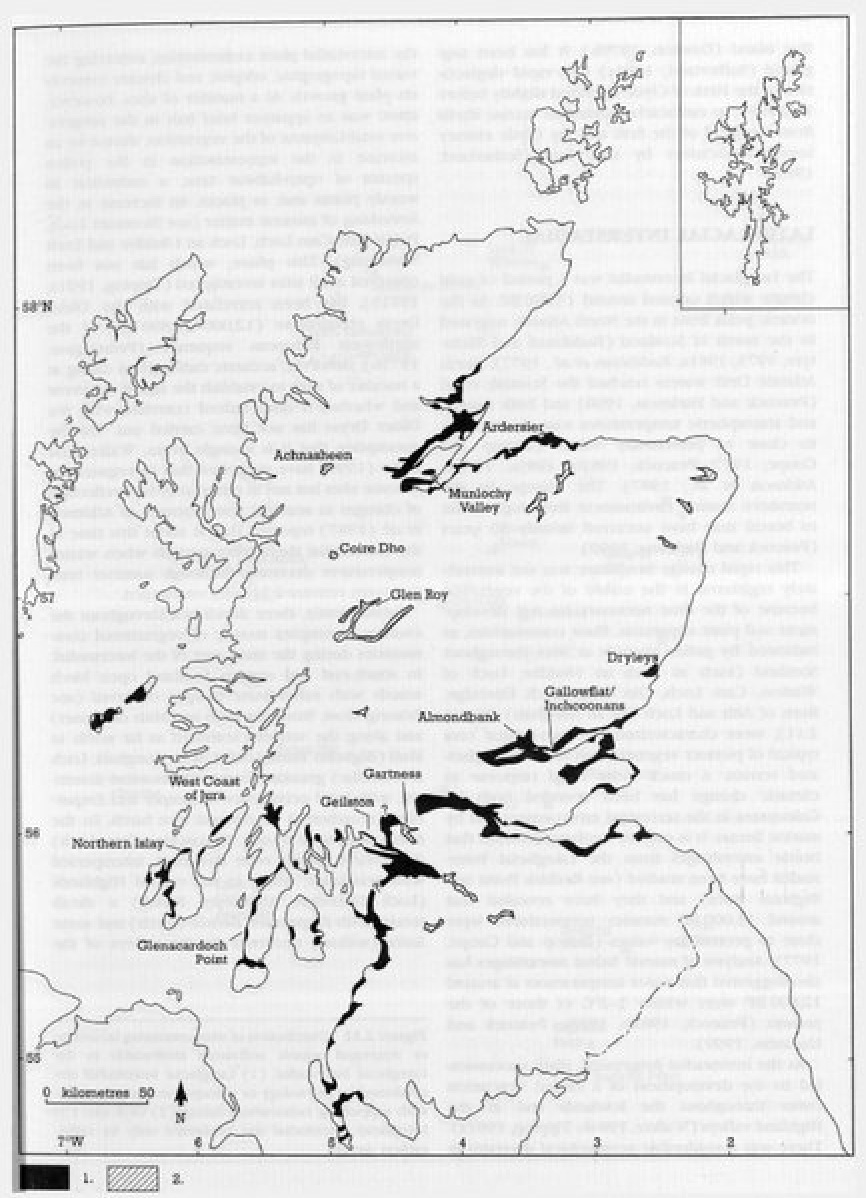
(Figure 2.7) Summary of the principal glacial, periglacial, marine and terrestrial depositional events recognized in the Quaternary record in Scotland. Note that the time-scale is not linear. Only those events that can be related to specific deposits on land or on the adjacent continental shelves are plotted.



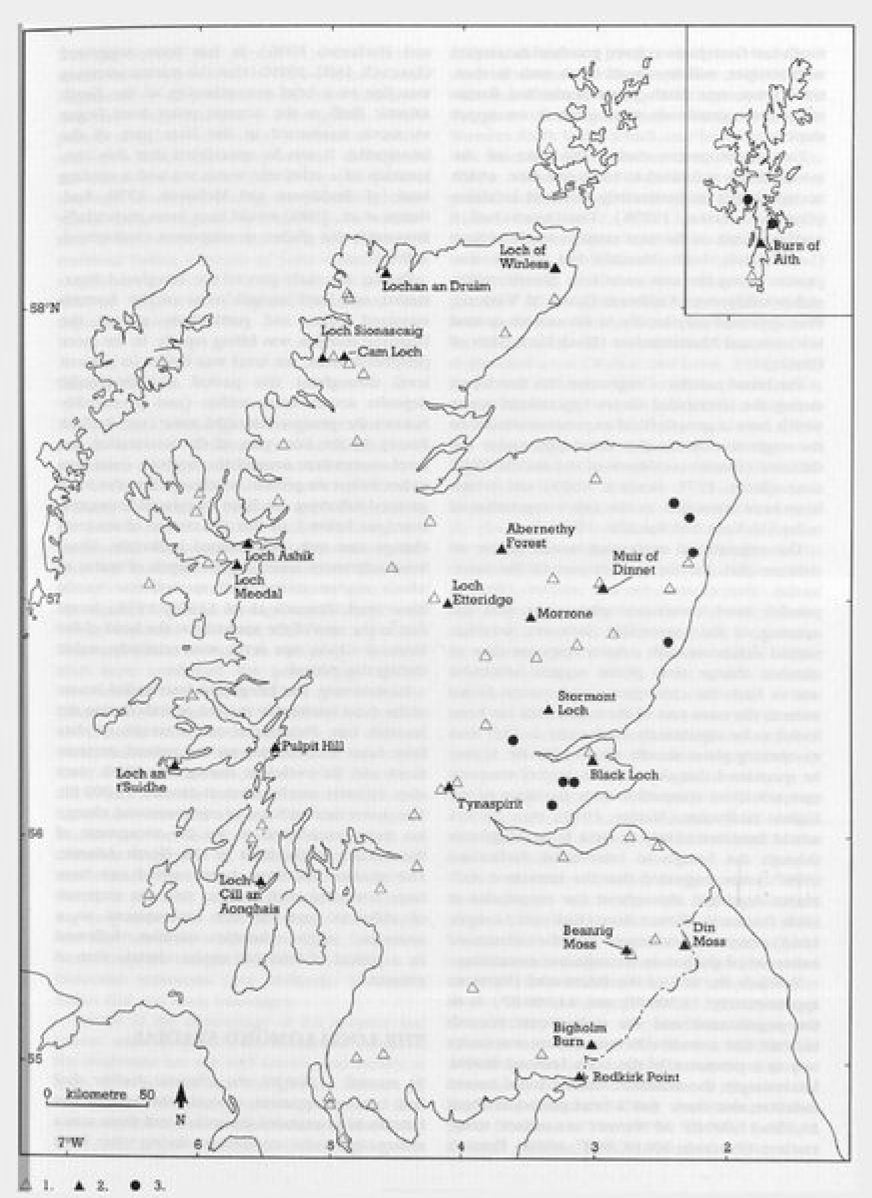
(Figure 2.8) Ice-sheet flow patterns and Loch Lomond Readvance glaciers.



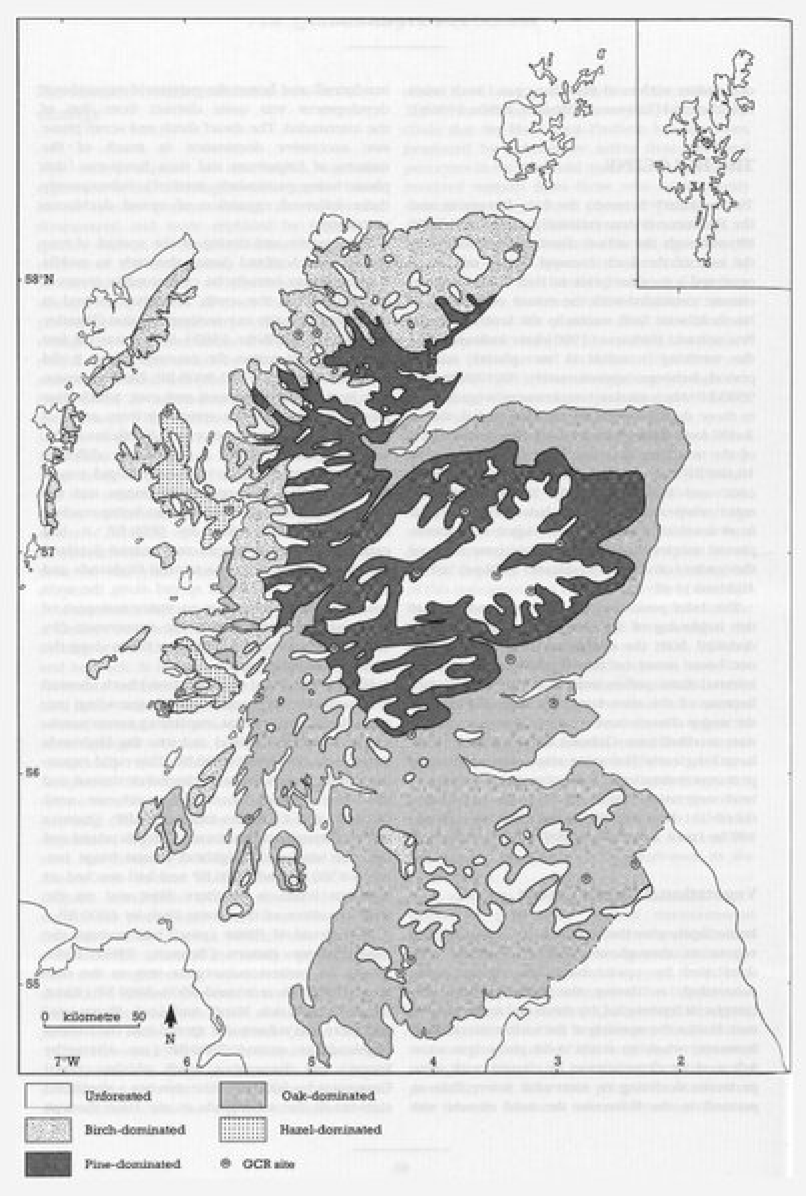
(Figure 2.9) Distribution of glaciofluvial deposits.



(Figure 2.10) Areas flooded by the sea during retreat of the Late Devensian ice-sheet (1), and location of ice-dammed lakes formed during deglaciation of the Late Devensian ice-sheet and the Loch Lomond Readvance (2).



(Figure 2.11) Distribution of sites containing lacustrine or terrestrial organic sediments attributable to the Lateglacial Interstadial. (1) Lateglacial Interstadial site confirmed by palynology or coleopteran analyses, many with supporting radiocarbon dating; (2) GCR site; (3) Lateglacial Interstadial site confirmed only by radiocarbon dating.



(Figure 2.12) Forest zones at approximately 5000 BP (from Bennett, 1989), and GCR sites with palynological data for the Holocene.